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Isolation basins, sea-level changes and the Holocene history of the Greenland Ice Sheet

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ABSTRACT

Isolation basins are natural topographic depressions that at various times in their history may be connected to or isolated from the sea by changes in relative sea-level (RSL). They provide a valuable source of data for tracking large scale (tens of meters), millennial-scale changes in RSL, as well as quiet-water depositional environments where abrupt changes caused by tsunami, iceberg roll or storms may be recorded. In this paper we review isolation basins as sources of RSL data with a particular focus on their use in Greenland to constrain the Holocene history of the ice sheet. A new RSL curve from Disko Fjord, West Greenland is presented, which shows that local ice free conditions were established at c. 11 k cal. yr BP, after which RSL fell rapidly from a marine limit at c. 80 m to reach close to present sea-level by c. 4 k cal yr BP. We compare this record with other isolation basin RSL data from six other sites in Disko Bugt and note a strong northwest/southeast differential rebound across the area during the early and mid-Holocene that reflects variations in ice load history. We compare the Disko Bugt data with other previously published isolation basin RSL records from Sisimiut (central West Greenland), Nanortalik (south Greenland) and Ammassalik (southeast Greenland). RSL fell below present during the early-Holocene at Nanortalik (c. 10 k cal yr BP) and during the mid to late Holocene elsewhere before rising to present. These differences reflect variations in the timing and amount of Greenland ice load change since the last glacial maximum, as well as non-Greenland processes, notably the collapse of the Laurentide forebulge and changes in ice equivalent sea-level. Isolation basin data have relatively small age and height uncertainties compared with other RSL indicators, enabling them to

resolve between different earth and ice sheet models, especially during periods of large ice load and RSL change.

Keywords: isolation basin, relative sea-level, Greenland Ice Sheet, isostasy, glacio-isostatic adjustment.

1. Introduction and aim

Isolation basins are natural depressions that at different times are either connected to or isolated from the sea by changes in relative sea-level (RSL). The basin sill controls the frequency of tidal inundation and defines the elevation of former RSL, whilst radiocarbon dating of basin deposits provides dating control. Staircases of isolation basins are an important source of RSL data to reconstruct glacio-isostatic adjustment (GIA), especially from sites that have experienced large scale (tens to hundreds of metres) of RSL change (Figure 1). Their use originated in the glacially-scoured landscapes of Sweden (e.g. Sundelin, 1917) where abundant lakes exist below and above the marine limit (the highest occurrence of marine deposits of postglacial age). They provide a vital source of RSL data here and also in Norway (e.g. Berglund, 1964; Fægri, 1944), Finland (e.g. Eronen, 1983), Northwest Russia (e.g. Corner et al., 1999), Scotland (e.g. Shennan et al., 1994; Shennan et al., 1999), Iceland (Lloyd et al., 2009) and Antarctica (e.g. Bentley et al., 2005; Cromer et al., 2005).

In contrast to these areas, isolation basins have only been studied in Greenland systematically for RSL research since the late 1990s, although their value was recognised earlier as demonstrated by the pioneering pollen and diatom work of Iverson (1953), Fredskild (1973, 1977, 1983) and Föged (1972, 1977). Instead RSL studies in Greenland have normally used radiocarbon dates from marine shells and driftwood in raised beaches or archaeological data (Funder, 1989; Funder and Hansen, 1996; Kelly, 1980; Rasch, 2000; Ten Brink and Weidick, 1975; Weidick, 1972). These data can have large age and height uncertainties (± 1 to 2 ka and ± 5 m or more) and because of their sometimes limited occurrence may be compiled from large geographical areas into composite RSL curves (e.g. Weidick, 1996). In contrast, the growing volume of isolation basin data from Greenland now provides strong local and regional constraints on the Holocene behaviour of RSL and hence the history of the Greenland Ice Sheet (GIS), and an important context for interpreting current and future ice sheet change (Long, 2009).

The aim of this paper is to review the use of isolation basins in RSL research with a focus on work from around the margin of the GIS. In Section 2 we introduce the isolation basin method and in Section 3 review previous RSL research in Greenland. In Section 4 we present a new isolation basin record from the Disko Bugt region of central West Greenland and combine this with other similar studies to reconstruct RSL and ice sheet history in this region through the Holocene. In Section 5 we draw together isolation basin studies from elsewhere in Greenland to explore regional patterns of RSL change and then look in detail at how this method can be used to constrain the mid to late Holocene RSL lowstand that occurs in many places around Greenland (Section 6). In Section 7 we explore the use of RSL data for developing GIA models of the GIS before reflecting on the challenges and opportunities for RSL research in Greenland in Section 8.

2. Isolation basins as indicators of relative sea-level change

Isolation basins vary in shape and size. Examples of very large basins include the Gulf of Carpentaria (Chivas et al., 2001; Reeves et al., 2008), the Baltic Sea (Andren et al., 2002; Björck, 1995; Eronen et al., 1990; Lampe, 2005), and the Black Sea (Ryan et al., 2003). Intermediate-sized basins include sea lochs and fjords, such as Loch Lomond in Scotland (Dickson et al., 1978) or Effingham Inlet on Vancouver Island, Canada (Dallimore et al., 2008). However, more common, and the focus of this paper, are relatively small lakes (typically <1 km^2 in area) that are most commonly used in RSL research.

There are three main phases of sedimentation in isolation basins that reflect stages in the isolation process associated with a RSL fall; i) a fully marine phase when the basin is continuously inundated by marine water; ii) a brackish-water, mixed salinity phase when the basin contains a mixture of marine and freshwater as the basin moves higher in the inter-tidal frame and the frequency of marine inundation falls, and; iii) a freshwater phase that follows basin isolation from the sea (Figure 1a). Under RSL rise a reverse sequence of sediments accompanies basin ingression. Basins may be isolated or ingressed several times and thus provide several sea-level index points.

2.1 The stratigraphy of isolation basins

Isolation basins used in RSL research are normally relatively small (<1 km^2) with water depths <10 m where the sample core is collected, usually using a hand-operated corer from a boat or raft. A problem with deeper water lakes is that they may contain old sea water trapped following basin isolation. Strom (1957, 1962) and Barland (1991) describe deep (>60 m) isolation basins in Norway that were isolated several thousand years ago and still today contain a c. 10 m-thick layer of saline water at their base. More preferable to study are smaller, shallow lakes that experience greater wind stress, ventilation through seasonal convection and a relatively larger input of freshwater from streams that together cause more rapid turnover and mixing of water (Snyder et al., 1997). These attributes mean that smaller basins provide a more direct record of RSL change compared to larger basins.

At the other end of the size spectrum, isolation basins that are too shallow can also be problematic because they may fill with sediment and cease to operate as an isolation basin, or after isolation may accumulate peat significantly above the level of the sill (e.g. Selby and Smith, 2007). This can restrict the ability of the basin to record a subsequent marine ingression. Shallow basins (<1 to 2 m water depth) may be affected by ice freezing to the lake bed that can disturb or stop sedimentation, whilst sea ice over a shallow basin sill can influence seasonal exchange of freshwater and marine water into a basin (e.g. Vestfold Mountains (East Antarctica) Cromer et al., 2005)).

Many studies rely on one or two sample cores collected from isolation basins (e.g. Bennike, 2000; Eronen et al., 2001; Long et al., 2003) but these risk not being representative of the overall basin stratigraphy. Completing stratigraphic transects across basins (e.g. Fægri, 1944; Kaland, 1984; Long et al., 1999; Shennan et al., 1994) allows mapping of the spatial variability in sediments which can reveal variable facies thicknesses that may be significant in subsequent laboratory analyses and interpretations, and can help identify areas of the lake that are less affected by sediment slumping or other disturbance (e.g. Linden et al., 2006; Romundset and Bondevik, 2011).

2.2 Determining sill elevation

Ideally a basin will have an exposed and intact bedrock sill that has not been eroded by stream outflow or human agency (Figure 1b). In well-vegetated landscapes, bedrock sills may be buried by peat requiring stratigraphic investigations to define the sill configuration (Corner et al., 1999; Shennan et al., 1996). Some studies assume that the present water level in the lake approximates the sill altitude, although Corner et al. (2001) note that water levels can be 1 to 4 m lower than the basin sill due to seepage through fractured bedrock. Sills may also be lowered artificially to aid basin drainage (Kaland, 1984). Thus, Eronen et al. (2001) describe lowering by ditches at basin thresholds of up to 2 m in Southwest Finland, or by road construction in Southwest Norway (Lohne et al., 2007). These sill elevation uncertainties require consideration within the context of overall RSL reconstruction. Where RSL fall has been very large, for example close to the centre of the Fennoscandinavian uplift, vertical sill uncertainties of several meters are less important than age uncertainties associated with dating basin isolation (Linden et al., 2006). However, sill height uncertainties become increasingly significant as the magnitude and rate of RSL change decreases.

2.3 The isolation / ingression contact

The most widely used sedimentology tools used for tracking basin isolation and ingression are loss on ignition and magnetic susceptibility (e.g. Linden et al., 2006; Lohne et al., 2007; Risberg et al., 1996; Sandgren et al., 1999; Snyder et al., 1997; Sparrenbom et al., 2006a; Sparrenbom et al., 2006b). Other, less commonly used analytical methods include pigments (Verleyen et al., 2004), alkenones (Bendle et al., 2009), bulk organic chemistry (Balascio et al., 2010; Mackie et al., 2005; Mackie et al., 2007), and scanning X-ray fluorescence (XRF) (Sparrenbom et al., 2006a).

A variety of biostratigraphic indicators of basin isolation are used either singly or by combining two or more methods, such as (e.g. Kaland, 1984; Kjemperud, 1981; Lie et al., 1983; Lohne et al., 2007; Long et al., 1999; Miousse et al., 2003; Shennan et al., 1994; Stabell, 1985), pollen (e.g. Hafsten, 1960; Shennan et al., 1999; Yu et al., 2003), plant macrofossils (e.g. Bennike et al., 2002; Björck et al., 1994; Lohne et al., 2007; Miousse et al., 2003; Solem et al., 1997; Sparrenbom et al., 2006a), foraminifera (Lloyd, 2000; Lloyd and Evans, 2002), ostracods (Sandgren et al., 1999), chironomids (Romundset and Bondevik, 2011) and dinoflagellate cysts (e.g. Kaland,1984; Shennan et al., 1999; Yu et al., 2003).

The most frequently used indicator of isolation is diatoms which are well suited to reconstructing the isolation process because different taxa have particular salinity preferences, meaning that distinct diatom assemblages can be used to characterise the marine, brackish-water and freshwater phases of basin isolation. Several authors record an abrupt increase in *Fragilaria* associated with basin isolation (Eronen et al., 2001; Risberg et al., 1996), although Stabell (1985) argues that care should be taken when interpreting changing frequencies of this taxon since it can display peaks below, at and above the isolation contact. A pronounced increase in insect remains (e.g. Chironomidae larvae) is also characteristic of isolation boundaries (Romundset and Bondevik, 2011).

2.4 Stages in the isolation process

Kjemperud (1981, 1986) provides a conceptual framework for interpreting changes in diatom assemblages from isolation basins and suggests four 'contacts' (Figure 1a, b):

The *diatomological isolation contact*: the sediment-water interface when the water in the photic zone of the basin became fresh. Kjemperud (1986) uses this term synonymously with the term 'isolation contact'.

The *hydrological isolation contact*: the final cessation of marine incursions into the basin. This is often coincidental with the diatomological contact in well mixed basins and may mark the transition to fully freshwater sedimentation.

The *sedimentological isolation contact*: the change from a minerogenic, mainly allochthonous sediment to a more organic, autocthonous deposit that formed within the lake basin. This contact is commonly observed below the diatomological/hydrological isolation contact and records when the sea ceases to transport significant amounts of minerogenic sediment into the basin. It may coincide with the development of anoxic bottom waters in a basin and the deposition of a finely laminated olive brown or black clay gyttja.

The *sediment/freshwater contact*: A horizon that defines the sediment-water interface at the time when any residual marine water is flushed from the basin. This only occurs in deep lakes that may hold trapped sea water (Strom, 1957).

Radiocarbon dates from isolation basins close to the local marine limit can provide limiting ages on the timing of local ice margin retreat. However, ages from the isolation contact provide only minimum dates for local ice free conditions because basal marine sediments may pre-date the isolation contact by several centuries or millennia. On the Kola Peninsula, Northwest Russia, Snyder et al. (1997) record pre-isolation dates that are up to 2 ka older than the isolation contact. Likewise, Romundset and Bondevik (2011) date molluscs in marine sediments from basins on Finnmark, northern Norway, that pre-date basin isolation by up to 4 ka.

As RSL falls and the sill enters the intertidal zone, some basins may experience an increase in tidal energy where water is channelled along bedrock troughs that connect to the open sea (see examples in Kaland (1984)). In other cases, the reduction in wave and tidal flow that accompanies isolation is associated with a fining-upwards sequence and an increase in organic content as basin productivity rises. As isolation approaches, so the frequency of marine inundation falls and a distinct brackish water phase occurs. This may be associated with the deposition of a finely laminated gytjja with variable quantities of minerogenic matter that is frequently olive grey to black in colour (e.g. Föged, 1972). This facies is interpreted to have formed in anoxic bottom water conditions with associated sulphur reduction, with occasional overtopping of saline, nutrient-rich water into the basin (Figure 1b). The thickness of this distinct facies varies depending on the basin properties and also the rate of prevailing RSL change. In rapidly uplifting areas or where basins are relatively small this facies may be thin or lacking altogether. When the rate of RSL change is low and the basin large and deep, the facies may be several decimetres or even metres thick. In Loch nan Corr, NW Scotland, Lloyd (2000) describes a gradual removal of marine conditions from fully marine to freshwater conditions over a period of c. 7000 cal yr that is recorded by c. 7 m of sediment.

There are conflicting ideas regarding whether the base or top of the laminated facies records the final basin isolation from the sea. Most authors suggest it is the top of this unit (e.g. Kaland, 1984; Kjemperud, 1981, 1986; Stabell, 1980; Svendsen and Mangerud, 1990). However, Corner and Haugane (1993) suggest that the laminated facies forms after final basin isolation, when salt water is still resident in the bottom of a basin (see also Corner et al., 1999). This difference of interpretation matters little where the facies is only a few centimetres thick and sedimentation is rapid, perhaps because of rapid RSL fall. Indeed, Eronen et al. (2001) date paired samples of brackish-water sediment and fresh-water gyttja in 29 isolation basins from Finland and conclude that the time difference between the two dates rarely exceeds 100 to 200 years, close to the age error associated with any single date. However, where the laminated facies is thicker, the age difference could be significant; a lag of several centuries (and metres of RSL change) may exist between basin isolation and start of the fully freshwater phase.

Following isolation, basin stratigraphies are dominated by freshwater lake sediment accumulation, normally comprising various mixtures of gytjja, plant macrofossils and mineral matter supplied from the local catchment. Although no longer flooded by tides or even storms, they can record exceptional marine events, such as large waves generated by iceberg roll, landslides or tsunami (Section 2.8).

2.5 Basin ingression by relative sea-level rise

The complex nature of RSL change in near- and intermediate-field locations mean that isolation basins may record one or more marine ingressions and isolations. The best known examples are the Holocene "Tapes" transgression in Norway (Corner et al., 1999; Svendsen and Mangerud, 1987), and mid to late Holocene transgressions observed in Scotland (Selby et al., 2000; Shennan et al., 1994). Other examples include an oscillation in RSL during the Younger Dryas in SW Norway (Lohne et al., 2004), and an early-Holocene transgression in Lake Ace, Vestfold Hills (Antarctica) (Cromer et al., 2005). There is extensive evidence for mid and late Holocene RSL rise in Greenland that we consider in Section 6.

2.6 Developing chronologies

There are several strategies for dating isolation basin sediments that are designed to minimise potential errors due to marine and freshwater reservoir effects and contamination by older or younger carbon. Initial studies focussed on bulk dating of freshwater sediment deposited immediately after basin isolation to avoid potential marine reservoir effects (Mangerud and Gulliksen, 1975). As a result these dates are slightly younger (or older) than the date for the isolation (or ingression) itself (Kaland, 1984; Mangerud, 1972). Contamination by younger carbon may be due to root penetration of aquatic macrophytes. In Western Norway, Kaland et al. (1984) and Kaland (1984) argue that penetrating roots of *Isoetes*, which grows in clear-water lakes up to 15 m deep, can result in 'young' ages for the isolation contact when bulk sediment is dated.

Following the advent of AMS ^{14}C dating, several studies have used marine molluscs (fragments or whole specimens, preferably of filter feeders in equilibrium with marine water) or discrete layers of marine brown algae and seaweed to date basal marine sediments and identify the timing of isolation (Romundset et al., 2009). During the brackish-water phase, choosing material such as terrestrial plant macrofossils that do not require reservoir correction is preferable, although Corner et al. (1999) use a variable reservoir correction for brackish bulk samples by subtracting a percentage of the marine reservoir age equivalent to the percentage of marine/brackish diatoms in the sample. The most robust approach is to develop a sequence of dates on terrestrial plant remains to estimate the age of the contact either through interpolation from age-depth curves (e.g. Linden et al., 2006; Risberg et al., 1996) or by Bayesian age modelling (Romundset et al., 2009). An excellent example of this approach is provided by Lohne et al. (2007) who date plant macrofossils from 12 contiguous 1 cm intervals across a short-lived brackish-water phase from a basin on Sekkingstadtjønn, Sotra (SW Norway) to precisely constrain the age of a sea-level lowstand during the Alleröd.

2.7 Reference tidal levels

A date from an isolation or ingression contact must be attributed to a particular tidal level (e.g. mean high tide (Kjemperud, 1981; Lohne et al., 2004; Romundset et al., 2009; Snyder et al., 1997) or highest astronomical tide (e.g. Sparrenbom et al., 2006a, b)). However different parts of an isolation sequence have the potential to provide several sea-level index points, each related to the same sill elevation but with a different reference tidal level. For example, Shennan et al. (1995) use diatom assemblages to develop index points from different phases of the isolation sequence, with reference tidal levels ranging from between mean low water of neap to just below mean high water of spring tides (a range of c. 2.5 m). This approach is particularly useful during slow RSL fall when basins may take several centuries or millennia to be isolated (Lloyd, 2000).

2.8 Exceptional events

Isolation basins from Norway and the Shetland Islands (Scotland) provide striking stratigraphic evidence for the tsunami that accompanied the Storegga submarine landslide, dated to c. 8200-8000 cal yr BP. This tsunami eroded lake bottoms and deposited graded and/or massive beds of sand, rip-up clasts, and coarse plant material in isolated basins close to high tide level (Bondevik et al., 1997a; 1997b, 1998). By coring staircases of lakes, the run-up of the Storegga-generated tsunami has been reconstructed to 10 to 12 m on the Western Norwegian coast (Bondevik et al., 1997a; Svendsen and Mangerud, 1990), 3 to 4 m in Finnmark, Northern Norway (Romundset and Bondevik, 2011) and, in combination with other stratigraphic evidence, up to 25 m on the eastern coast of Scotland (Dawson et al., 1990; Grauert et al., 2001; Smith et al., 2004).

In East Greenland, evidence for the Storegga tsunami is reported by Wagner et al. (2007) from Loon Lake (72° 53.2' N, 022° 08.3' W). At the time of the Storegga landslide, Loon Lake was a tidal inlet, 15 to 35 m deep and deposition was characterised by fine grained homogeneous sediments. These sediments are interrupted by a 0.72 m thick sandy horizon which has an eroded base and is associated with marked fluctuations in grain-size data and other physical and biogeochemical properties. Radiocarbon ages date this event to 8500-8300 cal yr BP which is close to that of the Storegga landslide. Long et al. (2008) also observe a distinct coarse unit within otherwise fine-grained sediments of marine origin in the base of basin QT7, near Ammassalik in Southeast Greenland (see Figure 2e of Long et al. (2008)). A radiocarbon date from the overlying isolation contact indicates that this coarse unit was deposited before c. 8160-7950 cal yr BP and might also possibly record disturbance associated with the Storegga tsunami.

Iceberg roll is another potential source of large waves entering a basin. For example, in the outer part of Disko Bugt on the west coast of Greenland, Long et al. (2003) detail stratigraphic evidence from the freshwater phase of a lake shortly after isolation and dated to c. 6000 to 5500 cal yr BP. The incursion was accompanied by some erosion of the underlying freshwater gytjja and the deposition of a thin (1 to 4 cm-thick) sheet of sand across the lake bed over a distance of c. 200 m. This may be evidence of iceberg roll or landsliding activity in the region at the time.

3. Relative sea-level changes in Greenland

For over 40 years, established approaches to RSL research in Greenland have focused on dating marine molluscs, driftwood, whalebone and archaeological material from raised beaches and deltas (Donner, 1978; Donner and Jungner, 1975; Funder, 1989; Hall et al., 2008; 2010; Ingólfsson et al., 1990; Kelly, 1985; Lasca, 1966; Rasch, 2000; Rasch and Jensen, 1997; Washburn and Stuiver, 1962; Weidick, 1972, 1976). These provide important constraints on the age of the marine limit and the trend in early and mid-Holocene RSL fall, which records GIA following glacial unloading and the retreat of the GIS since the LGM (Funder and Hansen, 1996). Sedimentary data from drowned beaches, archaeological remains and isolation basins (Kelly, 1980; Kuijpers et al., 1999; Long et al., 1999; Mikkelsen et al., 2008; Petersen and Hoch, 2004; Rasch and Jensen, 1997; Sparrenbom et al., 2006a) provide evidence for a switch from RSL fall to rise in the late Holocene that Kelly (1985) termed the 'Sandnes transgression' after a church in Ameralik (West Greenland) that, it is claimed, has been lowered by between 5 and 6 m since AD 1600 by crustal subsidence (Roussell, 1941).

Data is often compiled from wide geographical areas to develop a composite RSL curve for a region (Hjört, 1981; Ten Brink, 1974; Ten Brink and Weidick, 1975; Weidick, 1968; Weidick, 1996). In West Greenland this practice led Kelly (1985, p. 467) to comment "in no area do the data really justify expressing the emergence by a single well-defined curve; although some of the uncertainty may arise from the need to include data over relatively large areas, most of it is inherent in the data themselves". Kelly's (1985) concerns regarding the uncertainties in the data relate to the difficulties in ascribing a reference tidal level to the dated shell material in these studies and to problems in their age determination. For example, in a study of the mollusca fauna of Disko Bugt, Donner and Jungner (1975) dug sections in marine sands and gravels to a depth of 85 cm and sieved 100 to 200 g of shell material to obtain enough material for bulk radiocarbon dating. This led to an inevitable mixing of shells fragments of different ages. They report that different shell species from the same beach varied in age by up to 2400 ^{14}C years.

These uncertainties can be overcome by using AMS dating to target individual molluscs (or fragments), combined with the systematic definition of the altitude relationship between molluscs and their former sea-level. Thus, at Kjove Land and Schubert Dal, east Greenland, Hall et al. (2008; 2010) develop local RSL curves by dividing their ^{14}C dates into three types of index point: i) shell dates from fjord bed settings that provide maximum limiting dates on RSL; ii) shell dates from samples that can be fixed relative to exposed topset/foreset contacts in stratigraphic section which provide precise estimates of former sea-level, and; iii) shell dates that have an uncertain height relationship to former sea level. The main imitation of the mollusc-based approach is that it cannot track RSL if it fell below present, which in many parts of Greenland happened during the mid to late Holocene.

The many thousands of lakes below the marine limit in ice free areas of Greenland provide excellent opportunities for isolation basin RSL studies, with target lakes that vary in size from small, nearly-infilled basins a few tens of meters across, such as basin D4 from Disko Island (Section 4.3), to much larger tidal inlets such as IV2, a 3 km-long inlet at Innaarsuit, on the south shores of Disko Bugt (Long et al., 2003). Beyond their sheer abundance, the Greenland basins have several other attributes that make them suitable for isolation basin studies. Often their sills are exposed and can be surveyed directly, avoiding the need to probe deep peat sequences in an effort to define the sill altitude and geometry. Also, the vast majority of basins have been unaffected by significant human activity, such as inlet lowering for drainage or significant catchment modification by farming activities. There are though, some processes that can complicate their study. The harsh climate of the region means that rock faces surrounding basins can be destabilised and may collapse, potentially disturbing lake stratigraphic records or even impounding lakes sometime after RSL fell below the lake sill (Long et al., 1999). Frost shattering of bedrock sills can also complicate efforts to determine sill elevations, whilst lake ice can also modify lake shorelines, including their sills, by reworking and transporting sediment across them. Finally, iceberg roll or tsunami generated by landslide collapse (Section 2.8) provide rare but potentially important events that can disturb the sediment in isolation basins.

The limited vegetation cover and gneissic rock over much of West Greenland provides good field conditions to identify the marine limit, often defined by the lower limit of perched boulders or the upper limit of wave washed bedrock. It is rare to find datable shells or isolation basins close to the marine limit but because the rate of early and mid-Holocene RSL fall was so rapid, it is possible to infer the age of the marine limit – and hence when local ice free conditions became established – by upwards extrapolation of the shell-based RSL curve to intersect the marine limit (Weidick, 1972; Kelly, 1985; Hall et al., 2008). This assumes that RSL fell immediately after deglaciation, which may not always have been the case (Funder and Hansen, 1996).

A map of the marine limit for Greenland (Figure 2) shows three domes, over the Sisimiut district in West Greenland (maximum c.140 m), the Scoresby Sund region in east Greenland (maximum c.135 m), and on Hall Land in north Greenland (maximum c.130 m) (Weidick and Bennike, 2007). Low marine limits (c. 20 m) occur in regions where the ice sheet is currently close to the coast, such as in Melville Bugt in Northwest Greenland and in parts of Southeast Greenland. Where there is a wide ice free strip of land between the ice sheet and coast the limit has a domed appearance, rising from the outer coast to a maximum before decreasing towards the ice sheet margin (Funder, 1989; Funder and Hansen, 1996; Weidick, 1972, 1976). The marine limit around Greenland is diachronous and because it is a measure of changes in relative sea-level, its elevation cannot be used to directly infer changes in ice load.

The early- and mid-Holocene was characterised by rapid RSL fall as ice retreated from the continental shelf, in many areas reaching close to the present coastline at c.10 k cal yr BP and then retreating to a position inland of the present margin by the end of the Holocene thermal optimum (c. 5 k cal yr BP) (Funder and Hansen,1996). Rates of RSL fall in West Greenland are typically 10 m to 30 m/ka with many areas experiencing a slowing in the rate of fall during the mid-Holocene as the rate of glacio-isostatic rebound fell (e.g. Hall et al., 2008; Ingólfsson et al., 1990; Kelly, 1985). Only in one study in Northeast Greenland is there a suggestion of a mid-Holocene transgression which is based mainly on geomorphological evidence (Hjört, 1981).

Most RSL data from Greenland are from this period of early-Holocene rapid RSL fall, reflecting the availability of material for dating. However, steeply falling RSL curves in the early-Holocene suggest that in many areas RSL fell below present in the mid-Holocene before rising to present. There are two hypotheses, not necessarily mutually exclusive, for the cause of this late-Holocene RSL rise. Rasch and Nielsen (1995) argue that J-shaped RSL curves (proposed for Disko Bugt) reflect the collapse of the Greenland Ice Sheet forebulge. In contrast, Kelly (1980) and Weidick (1993; 1996) attribute late Holocene RSL rise to crustal reloading by neoglacial ice sheet expansion from its mid-Holocene minimum.

4. Relative sea-level changes in Disko Bugt, West Greenland

4.1 The Disko Bugt area

Disko Bugt is a large marine embayment located between 68º 30’ N and 69º 15’ N and 50º 00’ W and 54º 00’ W (Figure 3a). The bedrock geology of the mainland is Precambrian gneisses (Escher et al., 1976) with Teriary basalts dominant on Disko Island and the Nuussuaq peninsula. The western limit of the bay is defined by a bedrock ridge of gneiss that outcrops as a few small islands across the bay entrance (Figure 3b). In the south and west the landscape is heavily ice-scoured whereas alpine landscapes occur on Disko Island and Nuussuaq (Sugden, 1974). At the Last Glacial Maximum (LGM) Disko Bugt was over-ridden by ice fed by a number of ice streams, the largest of which today is Jakobshavn Isbrae (Roberts and Long, 2005). The ice sheet extended onto the continental shelf and remained there until the early part of the Holocene, before retreating rapidly across the bay to reach the eastern coast by c.10 to 9 k cal yr BP (Long and Roberts, 2002; Weidick and Bennike, 2007).

Disko Bugt is an important area for RSL research due to the well preserved mollusca faunas on the south shores of the bay where extensive glaciomarine deposits occur (Harder et al., 1949; Laursen, 1950). Donner and Jungner (1975) developed a composite RSL curve for the south of the bay using marine molluscs collected from a 100 km long west to east transect. They noted that the radiocarbon dates show a regular decrease in age with elevation, but they were unable to resolve a detailed RSL record or discern any differential patterns in RSL across the region.

Other RSL curves based largely on mollusc chronologies are presented from Disko Island by Donner (1978), Ingólfsson et al. (1990) and Rasch and Neilsen (1995). Ingólfsson et al. (1990) noted that the marine limit climbs across Disko Island from c. 60 m in the northwest to 90 m in the southeast (see also Rasch (2000)), suggesting that this reflects differential emergence and ice loading across the region as well as diachroneity in ice margin retreat. A combination of archaeological and morphostratigraphic studies (Rasch and Jensen, 1997; Rasch and Nielsen, 1995) all show that RSL fell below present during the late-Holocene but the magnitude and timing of the lowstand is not certain.

4.2 Isolation basin studies in Disko Bugt

The first systematic analysis of isolation basins in Disko Bugt to develop a local RSL curve is provided by Long et al. (1999) based on an analysis of nine isolation basins from Arveprinsen Ejland (Figure 3b). The resulting J-shaped RSL curve shows RSL fell from c.70 m at 9.9 k cal yr BP to reach a minimum of c. -5 m at 2.8 k cal yr BP before rising by an equivalent amount to present. Following this first study, further investigations were undertaken at four other locations in Disko Bugt to define the spatial variability in RSL across the study area. A transect of sites along the south coast, from

Qeqertarsuatsiaq in the west (Long and Roberts, 2003) through Inaarssuit (Long et al., 2003) and Akuliit and Nuuk (Long and Roberts, 2002) to Upernivik in the east was designed to construct a network of sites across the south and north of Disko Bugt (Figure 3b). To complete this original research aim we now describe a new RSL record from Disko Island below.

4.3 A new relative sea-level record from Disko Fjord, Disko Island

Much of the landscape of Disko Island is high relief and unsuitable for isolation basin studies. However, the Precambrian basement outcrops in several locations on the south of the island and also towards its centre in Disko Fjord, providing potential isolation basin targets (Figure 3b). Bennike (1995) previously cored two basins on the Qivittut peninsula at 16.5 m asl and dated the isolation contact in one to 4730±115 (5716-5056 cal yr BP). We now detail four new isolation basins located on or close to a neck of land at the head of Kangerluarsuk, a small tributary valley off Disko Fjord (Figure 3c). Donner (1978) suggests that west facing fjords on Disko Island, including the study site, were outlets of extensive valley glaciers at the LGM and that their outer parts became ice free at c. 11.5 k cal yr BP, with the marine limit in Disko Fjord at 85 m reached at c. 10 k cal yr BP. A single survey point suggests that the local marine limit close to the study area is c. 70 m asl (Donner, 1978; Rasch, 2000).

4.3.1 Methods

We cored four lakes from a small tethered boat using a Russian-type corer to map the stratigraphy of the sample lakes and collected sample cores for laboratory analysis using a Beeker corer. We surveyed sill elevations to a common benchmark that was in turn related to mean tide level. We also cored a tidal inlet with a sill at c. -2 m MTL but failed to identify any freshwater sediments and so did not collect a sample core. We analysed diatoms across the isolation contacts to identify levels of AMS radiocarbon dating of 1 cm thick bulk sediment samples that we calibrate using the Oxcal program (Bronk Ramsey, 2009) (Table 1).

4.3.2 Results

The highest basin sampled, D4, is a small roughly circular lake that appears to be a remnant of a once larger water body that has become partly infilled with sediment. It has a sill altitude of 80.77 m MSL and is located < 5 m below the local marine limit which is defined here by the lower limit of perched boulders. The open water in the basin extends over an area of c. 100 m^2 and a sequence of four cores at 5 m intervals revealed an exceptionally thick (c. 5 m) sequence of gytjja that overlies a grey silt sand (most postglacial gytjja thicknesses are < 2 m in West Greenland) (Figure 4, 5a). Diatoms from a sample core collected from station 2 shows a replacement of marine by brackish and then freshwater sediments, with the isolation contact dated to 9700±100 BP (11265-10734 cal yr BP).

Basin D3 has a sill altitude of 27.81 m MSL. We completed a transect of five cores at either 10 m or 5 m sampling intervals (Figure 4). The lithostratigraphy comprises a 1 to 2 m thick gytjja that overlies grey silts and sands. Diatoms from the sample core record basin isolation with a change from marine to freshwater conditions, and the isolation contact is dated to 7590±50 BP (8537-8323 cal yr BP) (Figure 5b).

D2 has a sill altitude of 14.68 m MTL. The lake is relatively shallow, with present water depths of c. 2 m, and contains a c. 1 m thick sequence of gytjja that overlies grey silt sands (Figure 4). Diatom preservation in the silt sands is poor but show that these minerogenic sediments are marine in origin and that the base of the gyttja formed under freshwater conditions. A date from the isolation contact yielded an age of 5580±40 BP (6439-6295 cal yr BP) (Figure 5c).

D1 is the lowest lake we sampled for laboratory analysis, with a sill altitude of 3.18 m MTL. The largest of the lakes sampled, a transect of eight cores shows a unit of gyttja that overlies grey silts and sands that drape the former lake bed (Figure 4). Diatoms analysed from a sample core from station 7 tracks basin isolation and a radiocarbon date from the isolation contact yielded an age of 3830±40 BP (4409-4096 cal yr BP) (Figure 5d).

The four radiocarbon dates from the isolation contacts are plotted in Figure 6a, together with two previously reported dates on lake sediment collected from the nearby area (from Eqalunnguit, 6750±105 BP (K-3505) (Föged, 1989) and Qivittut, 4730±115 BP (K-5576) (Bennike 1995) (Figue 3b and Table 1)). The highest basin (D4), c. 5 m below the local marine limit, was isolated at c.11 k cal yr BP and this provides a minimum age for the establishment of local ice free conditions at contemporaneous sea level. This is the oldest date reported from the fjords of western Disko Island and suggests that within a few centuries of the opening of the Holocene the glaciers of the main Disko Island ice cap had retreated to at least the mid part of Disko Fjord. The date also provides a minimum age for the establishment of open marine conditions on the southwest coastline of Disko Island, suggesting Disko Fjord ice was separate from the main Greenland ice sheet by this time.

The initial fall in RSL was rapid, from c. 80 m at 11 k cal yr BP to c. 25 m at 8.5 k cal yr BP at a rate of >20 mm yr. Although the basins are widely spaced, we suggest the rate of RSL then slowed and RSL reached close to present by c. 4 k cal yr BP.   The isolation contact date reported by Bennike (1995) from Qivittut (10 km south of our field area) plots above the data point from D3 and the trend depicted by the other data. We have no direct evidence that RSL fell below present in the late Holocene at our field site, although Donner (1975) and Rasch and Nielsen (1995) suggest a small transgression occurred on the south coast of Disko Island sometime after c. 4 k cal yr BP.

The RSL data from Disko Island and Arveprinsen Eijland (Long et al., 1999) in the north of the bay are broadly similar (Figure 6a), although the marine limit is higher and slightly older at our Disko Island site (80 m asl, c. 11 k cal yr BP) compared to Arveprinsen Eijland (70 m asl, c. 10 k cal yr BP). This reflects the earlier date for the establishment of ice free conditions at the former site. During the mid-Holocene RSL fell to present c. 4 to 5 k cal yr BP and, at Arveprinsen Eijland, continued to fall to a lowstand of c. -5 m at c. 3 k cal yr BP before rising to the present. During the mid-Holocene, two data points stand out as potential outliers from basin V8 on Arveprinsen Ejland (Long et al., 1999) and, as noted above, from Qivittut (Bennike, 1995) (Figure 6a) and deserve further consideration since they give the impression of a possible temporary mid-Holocene slow-down in RSL.

Previously Long et al. (1999) rejected the V8 isolation contact date, arguing that it was likely contaminated either during sampling or it reflected disturbance, perhaps by sediment slumping. The stratigraphy of the V8 basin exhibits a lower impenetrable red/brown, humified *turfa* that is overlain by marine silts and clays with an abrupt transition into an overlying freshwater gytjja that extends to lake bed. Sometimes a thin (2 mm) bed of sand occurs at the interface between the silts and the gytjja, and diatoms indicate an abrupt change to freshwater conditions with no intermediate brackish water phase. The bulk radiocarbon date was taken immediately above the contact between the marine silts and the freshwater gytjja, making it a minimum age for basin isolation. Overall, the stratigraphy suggests that the basin sediments were disturbed following isolation, with the lower contact of the sampled gytjja forming some time after basin isolation. In contrast to the V8 data point, the litho- and biostratigraphy of the Qivittut sample (Bennike,1995) record gradual changes in environment and no sign of any discontinuity, whilst the radiocarbon age conforms with two other dates higher in the same sample core.

We think it unlikely that the data from V8 and Qivittut record a real slow-down in mid-Holocene RSL. First, such a feature is absent from all other existing RSL records from Disko Bugt and indeed, at other sites in Southern Greenland. This suggests that the data record a local and not regional process, restricted to the northern part of Disko Bugt. Second, there is no obvious local (Greenland) glacio-isostatic mechanism that could cause a RSL slow-down in northern Disko Bugt during the mid-Holocene, when all available palaeoclimate data suggest that West Greenland was experiencing relatively warm conditions during the Holocene thermal optimum (Weidick and Bennike, 2007). Third, there is only one instance in the Disko Bugt isolation basins for basin ingression during the mid-Holocene, in basin IV4 from Innaarusit where a brief inwash event is recorded between c. 6000 to 5500 cal yr BP, but this incursion appears short-lived and associated with a storm surge, or a wave generated by a landslide or iceberg roll (Long et al., 2003). Finally, existing glacio-isostatic rebound models for Disko Bugt all predict a smoothly falling RSL during the early and mid-Holocene, with no evidence for a millennial-scale slow-down at the time suggested (Tarasov and Peltier, 2002; Simpson et al., 2009).

4.4 Comparisons with other Isolation basin relative sea-level studies in Disko Bugt

All isolation basin RSL data from Disko Bugt are plotted in Figure 6b where we combine the data from Innaarsuit and Qeqertarsuatsiaq because of their similarities. At all sites RSL fell quickly from the marine limit during the early-Holocene, slowed during the mid-Holocene and fell below present after c. 5 to 4 k cal yr BP. The highest isolation basin is from the most westerly sites (Qeqertarsuiatsiaq / Innaarsuit) and records RSL at c. 100 m asl at 10.4 k cal yr BP (Long and Roberts, 2003) and c. 108 m asl at 10.2 k cal yr BP (Long et al., 2003) respectively.

In the southeast of the bay the marine limit on East Akulliit is c. 85 m (Long and Roberts, 2003) and falls steeply in an easterly direction to c. 54 m asl at Nuuk over a distance of c. 10 km. This likely records a slow-down in ice margin retreat across this part of the bay (and a possible readvance during the ‘8.2 k yr event’, Long and Roberts (2002)) as the ice sheet became grounded on the higher terrain and fjord landscapes in the east of Disko Bugt (Warren and Hulton, 1990; Weidick, 1972). At Upernivik, mid-way along Jakobshavn Isfjord and close to the ice sheet margin, the marine limit is at c. 60 m asl and dates to c. 8 k cal yr BP (Long et al., 2006). One drowned tidal basin at this site shows RSL here fell below present c. 5 to 4 k cal yr BP to reach a lowstand at c. 2.5 to 3 k cal yr BP before rising 5 m or so to present.

4.4.1 The timing of ice margin retreat across Disko Bugt

Dates from seven isolation basin studies, lakes above the marine limit as well as the oldest available mollusc dates constrain the timing of ice margin retreat across Disko Bugt. Long et al. (2003) and Weidick and Benike (2007) provide syntheses of these data that we update with our new age estimate from Disko Fjord (Figure 7). The latter data point confirms that the northwest of Disko Bugt and Disko Island became ice free earlier than the main bay itself, with oldest dates of c. 12.4 and 11.6 k cal yr BP in the northwest of the study area, c. 11.0 k cal yr BP at Disko Fjord and c. 10.5 k cal yr BP at Qeqertarsuaq on the south of Disko Island. Ice free conditions are recorded in the western part of Disko Bugy by c. 10.3 k cal yr BP, indicating rapid gutting of ice from the bay during the early-Holocene.

4.4.2 Holocene relative sea-level in Disko Bugt

The largest spatial gradients in RSL in Disko Bugt occur during the early-Holocene, when RSL in the southeast was falling significantly faster than in the northwest. This pattern persists through the Holocene, albeit less pronounced after c. 6 k cal yr BP. During the late-Holocene, the lowstand and start of the late-Holocene RSL rise is recorded first in the east, at Upernivik and Arveprinsen Eijland (c. 3 k cal yr BP, no data exist from Akulliit and Nuuk) and later in the west (c. 2 k cal yr BP). This pattern is compatible with the progressive expansion of a belt of peripheral subsidence from the ice sheet due to regrowth during the neoglacial (Rasch and Jensen,1997; Long et al., 1999).

The isolation basin and other RSL data from Disko Bugt do not reveal obviously strong spatial or temporal differences, and in many instances the data from different areas are rather similar. This pattern is contrary to what one might expect, with shorelines being progressively tilted as one moves closer to the ice sheet margin (e.g. Weidick, 1996). There are several possible reasons for this. First, Disko Bugt was rapidly gutted of ice during the early-Holocene meaning that large spatial differences in ice unloading only happened in the eastern part of the bay, where the rate of ice margin retreat slowed and was interrupted by an early-Holocene readvance (Long and Roberts, 2002; Young et al., 2011). Second, the unloading of the Earth's crust by ice was, across Disko Bugt, partly replaced by a simultaneous re-loading by water as the bay was flooded by the sea. Water depths across Disko Bugt are presently 200 to 400 m (locally over 700 m), and would have been higher by up to 100 m or so at the start of the Holocene (Long and Roberts 2003). Thirdly, glacio-isostatic adjustment modelling (e.g. Simpson et al., 2008) and the RSL observations described above suggest a northwest to southeast RSL gradient across Disko Bugt, meaning that sites along the south shores of the bay, although separated by several tens of kilometres in a west to east direction, record broadly similar RSL histories.

4.5 Disturbance of lake basin stratigraphies

As discussed in Section 3, landslides or iceberg roll are potential sources of extreme waves into basins, especially during the early and mid-Holocene when calving outlet glaciers would have been numerous in the main part of the bay. Compared to the other areas studied in Disko Bugt, the lake stratigraphies on Arveprinsen Eijland (Long et al., 1999) are more disturbed at or about the time of basin isolation compared to other sites we have studied. The Arveprinsen Ejland basins are in a narrow fjord on the northwest coast of the island, which faces the Nuussuaq peninsula. A recent coastal landslide at Paatuut on this peninsula generated a wave which destroyed the village of Saqqaq, 40 km away (Figure 3b). The tsunami washed icebergs up to 800 m inland and up to 40 m asl (Dahl-Jensen et al., 2004; Pedersen et al., 2002).

Previous landslides along this coast have occurred at a rate of 1 to 2 per thousand years (Dahl-Jensen et al., 2004). There is therefore a possibility that at least some of the disturbances observed in the lake stratigraphies on Arveprinsen Eijland (and perhaps also elsewhere in Disko Bugt) record large waves generated by coastal landslides and / or iceberg roll. Studies from close to the calving margin of Jakobshavn Isbrae, (Amundson et al., 2010; Amundson et al., 2008) provide observational measurements of tsunami waves generated by iceberg roll close to the terminus that routinely have amplitudes of c. 2 m. In confined settings, such as within a fjord, iceberg roll can cause waves of a similar size. MacAyeal et al. (2011) suggest that as a rough rule of thumb, a tsunami with a height of 10 m could be generated by the capsize of an iceberg from Jakobshavn with a typical thickness of 1000 m.

5. Isolation basin relative sea-level records from elsewhere Greenland

We now compare the isolation basin RSL records from Disko Bugt with isolation basin records from three other parts of Greenland; Sisimiut (central West Greenland), Nanortalik (South Greenland) and Ammassalik (Southeast Greenland), to examine broad trends in RSL in southern Greenland.

5.1 Sisimiut (central West Greenland)

Sisimiut lies close to one of the main centres of postglacial uplift in Greenland where the marine limit reaches an elevation of c.140 m (Figure 2) (Funder,1989). At the LGM the ice sheet extended well onto the continental shelf, perhaps reaching as far as the shelf break (Bennike and Björck, 2002). This implies an overall retreat in this area of nearly 300 km since the LGM, probably more during the mid-Holocene when the ice sheet was smaller than present (Kelly,1985). Cosmogenic exposure ages from the coastal mountains suggest that the ice sheet had a maximum thickness here of 810 to 755 m and began thinning after c. 21 k yr BP (Rinterknect et al., 2009; Roberts et al., 2009). Isolation basins have been studied in two areas, close to the town of Sisimiut (Bennike et al., 2011) and about 40 km to the south (Long et al., 2009). The former targeted lakes close to the marine limit, whereas the latter focused on mid and late Holocene RSL changes only. Both studies are located on the outer coast and the data are plotted on a single graph in Figure 8, although it is noted that there may be some unaccounted differential crustal rebound between the two areas.

RSL fell rapidly from the marine limit which is dated here to shortly before c. 10.9 k cal yr BP based on the highest sample of the marine bivalve *Macoma calcarea* from the base a lake at c.100 m asl (Bennike et al., 2011). The initial rate of RSL fall was c.15 m per 100 yr, slowing after c. 9 k cal yr BP. There is only one basin between c.100 m and c.15 m asl, although radiocarbon dated marine shells from this height range indicate RSL continued to fall during this interval (Petersen and Hoch, 2004; Long et al., 2009). The lowest isolation basin of Bennike et al. (2011), at c. 18 m asl, is very shallow and contains an incomplete sedimentary record so is not included here. The data from Long et al. (2009) show that RSL fell from 14 m asl at c. 5.7 k cal yr BP to reach a lowstand of -4.0 m at c. 2.3 to 1.2 k cal yr BP before rising by an equivalent amount to present.

5.2 Nanortalik / Qaqortoq / Tasiusaq (South Greenland)

Isolation basin RSL records exist for the Nanortalik, Qaqortoq and Tasiusaq areas of South Greenland (Bennike et al., 2002; Fredh, 2008; Randsalu, 2008; Sparrenbom et al., 2006a; Sparrenbom et al., 2006b), providing detailed RSL reconstructions based on basins above and below present sea level (Figure 8). Located on the southern tip of Greenland, the outer coast here became ice free after c.15 k cal yr BP which is the earliest known date from Greenland (Bennike et al., 2002). Three separate RSL curves exist for the study area that show the same general pattern, although the curves from Qaqortoq and Tasiusaq are younger and show a steeper initial fall in RSL that reflects their closer position to the ice sheet margin. For ease of viewing we present data from only the oldest record from Nanortalik (Sparrenbom et al., 2006a) in Figure 8.

The marine limit in the area is below 55 m asl and isolation basins show that RSL fell rapidly during the Late Glacial and early-Holocene to reach a lowstand of perhaps -6 to -8 m before rising to present in the last 6 to 4 k cal yr BP (Sparrenbom et al., 2006a). The rise in RSL to present appears to have begun first at Nanortalik and more recently at Tasiusaq, close to the ice sheet margin, although there are difficulties in reconstructing a precise chronology for this part of the record. This is opposite to the pattern recorded in Disko Bugt where the late Holocene RSL rise occurs first at sites close to the ice margin. This suggests that the transgression here records a combination of effects caused by the collapse of the Laurentide forebulge and the neoglacial expansion of the ice sheet (Fredh, 2008; Randsalu, 2008; Weidick et al., 2004).

5.3 Ammassalik (Southeast Greenland)

There are almost no RSL data between the Scoresby Sund region of central East Greenland (Hall et al., 2008; 2010; Pedersen et al., 2011) and the southern tip of

Greenland. One isolation basin record is provided from the Ammassalik area of Southeast Greenland, c. 600 km north east of Nanortalik (Long et al., 2008). The marine limit here is c. 70 m asl and is dated to 11 k cal yr BP based on dates for the onset of organic sedimentation in lakes above the marine limit. This is in agreement with cosmogenic nuclide age determinations from the same area (Roberts et al., 2008). Four isolation basins show RSL fell to c. 24 m asl by c. 9.5 k cal yr BP and continued to fall at a decreasing rate thereafter to reach close to present by c. 6.5 k cal yr BP (Figure 8). Although basins below present sea-level were cored, Long et al. (2008) were unable to demonstrate any evidence that RSL fell below present although this must be a possibility based on the mid-Holocene trend.

5.4 Comparison of isolation basin records from Greenland

The records described above reveal clear regional differences in RSL change. In West Greenland, the RSL records from outer Disko Bugt (the combined Qeqertaruatsiaq / Innaarsuit record) and Sisimiut are very similar, with a rapid fall during the early and mid-Holocene, passing below present after c. 4 k cal yr BP and rising by up to 5 m in the last 2 to 3 k cal yr. There seems to be little differential crustal movements between these sites (Figure 8), despite the differences in the elevation of the marine limit (c. 110 m and 140 m respectively).

These J-shaped RSL records span the Holocene only, in contrast to the records from Southern Greenland that begin in the Late Glacial. The initial RSL fall at Nanortalik is as rapid as in West Greenland, but RSL fell below present significantly earlier (c. 10 to 8 k cal yr BP) and reached a lower low-stand (-6 m to -8 m) sometime in the mid-Holocene and certainly before the onset of neoglacial cooling which generally post-dates 4 k cal yr BP (Long et al., 2009). It is not known where the transition from a Nanortalik-type to a Sisimiut-type RSL record occurs in West Greenland because of an absence of RSL records from between these areas. However, using a small number of marine shells and lake basin data, Weidick et al. (2004) suggest that the mid-Holocene lowstand occurs progressively later as one moves north, such that at Paamiut the RSL record closely resembles that of central West Greenland with RSL only falling below present after 5 k cal yr BP. More data are required to test this hypothesis further. Lastly, the Ammassalik record spans the Holocene only and shows the characteristic J-shaped history observed at all sites, but with a rapid fall in RSL during the early-Holocene more similar to that observed in South Greenland rather than central West Greenland. The early-Holocene RSL fall in these areas indicate fast removal of a significant volume of ice during initial ice margin retreat (Sparrenbom et al., 2006a, b; Long et al., 2008).

6. Defining the Holocene relative sea-level lowstand in Greenland

Isolation basin studies provide important constraints on the Holocene RSL lowstand in South and West Greenland. However, the use of ingression contacts for RSL studies are often problematic because of uncertainties in determining their age, the sill elevation and also interpreting their biostratigraphic records. At Sisimiut, for example, two drowned basins with only slightly different elevations (<0.3 m) each indicate a short interval of largely freshwater conditions (although a persistent marine influence is present in one), dated to c. 3 and 2.2 k cal yr BP respectively. However, the radiocarbon dates suggest that the lower of the two basins was isolated and ingressed about a thousand years before the slightly higher basin. This may record problems in dating the basin ingressions, with some reworking of the soft lake sediments accompanying basin ingression. Similar problems are identified in Nanortalik (e.g. lake N29), where Sparrenbom et al. (2006b) record a hiatus caused by erosion during marine ingression.

Establishing the elevation and nature of contemporary tidal basin sills is also not easy. For recently ingressed basins, it may be possible to survey the sills directly at low tide and often there is a clearly defined channel that connects the open sea to the basin. However, sills can be covered in seaweed and the channel can have several bedrock steps within it. The sill may also have been eroded over time and thus lowered from its original elevation at the time of basin ingression. This is especially so for basins that exchange large volumes of water during rising and falling tides (Figure 9). As sill depths increase, so it becomes necessary to survey them from a boat at slack water on a low tide using an echo-sounder or survey staff. If the sill is boulder-covered, it is necessary to complete transects across the sill, making corrections for any changes in tide level during the survey, to determine the sill configuration. All these factors mean that it is difficult to establish the sill elevation of contemporary basins, especially those that are significantly below present. When the amount of RSL rise recorded by these basins is small, (e.g. less than 2 to 3 m), these uncertainties can introduce a significant error in the resulting RSL reconstructions.

Compared to isolation contacts, ingression contacts can be difficult to identify in sample cores. Where erosion is not evident, the lithological contact may be gradual, indicated by a subtle up-core increase in fine minerogenic sediment (e.g. Sparrenbom et al., 2006b). The microfossil distributions during the isolated phase are not always entirely fresh, with a lingering signal of brackish water and marine diatom taxa commonly recorded. This type of mixed assemblage can make the precise identification of the ingression contact difficult (for examples, see basins Nag-2 and Nag-1 in Long et al. (2009)).

The identification of fully fresh lake sediments in a basin that now lies several meters below present sea level provides unambiguous evidence that RSL has risen since their deposition. However, for the reasons discussed above, RSL chronologies provided by ingression contacts need treating with care and provide at best long-term trends in RSL with millennial, meter-scale chronological resolution. A critical challenge in their use is to identify basins with sills at appropriate elevations and alternative approaches to RSL reconstruction can provide higher quality data, notably salt marsh deposits that have accumulated in the last millennium (Long et al., in review; Long et al., 2010; Woodroffe and Long, 2009, 2010). These latter studies highlight a slow-down in the late Holocene RSL at sites in south and West Greenland that occurred sometime after AD 1600, demonstrating that linear interpolation of RSL from ingression contacts of late Holocene age to present should be avoided where possible (Long et al., 2010).

The late Holocene rise in RSL is often cited as evidence for a significant readvance of the Greenland Ice Sheet during the neoglacial (Kelly, 1985; Weidick, 1996). However, we caution against this interpretation because it is clear that RSL in much of Southern Greenland fell below present during the early- and mid-Holocene, and started to rise to present before the onset of the neoglacial climatic deterioration. Moreover, the significance of non-Greenland processes during the mid- and late-Holocene, especially the collapse of the peripheral bulge that surrounded the Laurentide Ice Sheet, means that attributing a Greenland ice load change as the main driver of the late Holocene RSL rise is not possible.

7. Using isolation basin data to constrain Greenland Ice Sheet models

Glacial isostatic adjustment models require good quality ice margin position and RSL data for calibration. In Greenland, the isolation basin data described above are preferred over other RSL data because of their smaller age and altitude uncertainties compared to other data sources that means that they can be used to discriminate between different ice and earth model parameterizations. In South Greenland, Bennike et al. (2002) use an ice sheet model with an initial radius of 400 km and a central ice thickness is 2800 m, which approximates an LGM ice cover extending to the shelf edge. Melting started at 15 k yr BP and continued until 10.5 k yr BP and then ceased. By varying the ice and earth components of this model, Bennike et al. (2002) demonstrate that the reduction in ice thickness over the coast must have been substantial (c. 1500 m) and also that the retreat from the outer shelf must have occurred rapidly, from c. 15 k yr BP. Because the model predictions include a significant mid-Holocene RSL highstand which is not observed in the Southern Greenland RSL records, the RSL history here must include a contribution from the melting of the nearby North American ice sheets and / or a late-Holocene readvance of the Greenland Ice Sheet (Bennike et al., 2002).

Fleming and Lambeck (2004) subsequently developed a GIA model for the entire GIS using RSL observations from all of Greenland, including isolation basin data from South Greenland and Disko Bugt. They compared their initial model to others that allow for neoglacial readvances of the GIS of 10 km and 20 km, starting in the mid- and late-Holocene. The inclusion of a neoglacial advance causes a late-Holocene RSL transgression. Using the Arveprinsens Ejland RSL record from Disko Bugt (Long et al., 1999), Fleming and Lambeck (2004) model a maximum RSL lowering of 6 m below present caused by a modelled neoglacial ice sheet readvance of c. 40 km, starting at 2.5 k yr BP.

In an alternative approach, Tarasov and Peltier (2002) develop a thermomechanical ice-sheet model (GrB) that they combine with the global GIA model ICE-5G (Peltier, 2004). They calibrate the model against RSL data from across Greenland but do not use isolation basin records, which at the time of analysis were limited in number. The scatter in the RSL data meant that it was not always possible to discriminate between earth and ice model parameterisations, including calving and temperature forcing. To fit model predictions to the observed rise in late Holocene RSL in West Greenland, Tarasov and Peltier (2002) start a neoglacial reloading of Greenland at 9 to 8 k yr BP, significantly before any glaciological or palaeoclimate data suggest such a change might have occurred.

In a regional modelling exercise in Southeast Greenland, Long et al. (2008) use Tarasov and Peltier’s (2002) GrB model and an alternative model (Hu1) tuned to fit observations of ice extent only (Huybrechts, 2002). In seeking to reconcile model predictions and isolation basin RSL observations, Long et al. (2008) fail to obtain a close fit and note this cannot be explained by modelled Earth viscosity structure alone, instead suggesting that this most likely reflects limitations in the two ice models considered. As with Bennike et al. (2002), the analysis finds that a thicker ice sheet model that extended well onto the shelf at the LGM provides an improved fit. This is compatible with research elsewhere that also supports an enlarged LGM ice sheet that extended onto the continental shelf in West and East Greenland (Evans et al., 2002; Hakansson et al., 2007; O Cofaigh et al., 2004; Roberts et al., 2009; Roberts et al., 2008; Wilken and Mienert, 2006).

In the most recent ice sheet modelling exercise, Simpson et al. (Simpson et al., 2009) tune a revised ice model (Huy2) to the isolation basin RSL data from Disko Bugt and elsewhere in Greenland, combining these constraints with other evidence of ice margin position since the LGM (Figure 10a). The main changes to the model include a more extensive LGM ice sheet, modifying the temperature forcing by including a Holocene thermal maximum, and changing the sea-level forcing parameter to produce a relatively late (12 k cal yr BP) and rapid ice sheet retreat. Where there are no isolation basin records, the large scatter in the mainly mollusc-based data means it is not always possible to discriminate between different model parameterizations. In West Greenland, the Huy2 model predicts RSL fell below present during the mid- to late-Holocene before rising to present and that the ice sheet here retreated inland of its present position by up to 80 km between 8 and 6 k yr BP before readvancing during the neoglacial (Figure 10a). The model / observation fit is generally best in West Greenland, where the main model tuning was undertaken, and fit in the south of Greenland and along the east coast is generally poorer, pointing to unresolved questions regarding the ice and Earth models in these areas.

7. Conclusions

Isolation basins are amongst the most valuable sources of sea level data from near- and intermediate-field locations and are especially useful in tracking large scale changes in sea-level (tens to hundreds of meters) over millennial-time scales. Their low energy depositional environments allow sediments to accumulate that record the different stages in lake isolation and ingression, including the impact of extreme events such as tsunami or waves generated by storms, iceberg roll or landslides. Where the local geology is suitable, staircases of basins in a small geographical area can be used to develop local RSL records that extend from above the local marine limit to present-day sea-level or below. The main limitations as RSL data points relate to uncertainties in defining the elevation of the sill and in overcoming disturbance of basin sediments, especially during ingression by gradual or sudden RSL change.

In Greenland, isolation basins have been studied for over 50 years but only recently have detailed RSL records started to be developed. The currently available data is skewed heavily to Disko Bugt and the southern tip of Greenland which between them account for the majority of available data. However, these studies have enabled important new developments in our understanding of the history of the Greenland Ice Sheet, by producing data that can track the timing of ice margin retreat and advance, as well as providing powerful constraints on glacio-isostatic adjustment models of the entire ice sheet (e.g. Tarasov and Peltier, 2002; Simpson et al., 2009). The improved age and altitude control compared to other methods of RSL reconstruction means such records are able to discriminate between some model paramaterisations that other data cannot. The geographical coverage of sites is expanding in West Greenland through time, with a new record recently published from the Sisimiut area (Long et al., 2009; Bennike et al., 2011) and a further record in preparation from Paamiut by the current authors. Future work should target other areas of Greenland, especially in the Southeast where the Ammassalik RSL record (Long et al., 2008) remains the only study for this entire sector of the ice sheet. In addition, isolation basins have the potential to provide important information on the risk posed by large waves generated by landslides or ice berg roll.

More generally, future work is required to better understand the controls on the process of isolation itself, including the specific depositional conditions associated with the accumulation of the brackish water deposit that separates the marine from freshwater phase. Important advances have been made in this regard through studies in Scotland of contemporary physical and ecological processes operating in basins at different stages of isolation (e.g. Lloyd, 2000; Lloyd and Evans, 2002; Mackie et al., 2006; Bendle et al., 2009). A simple hydrological model that includes parametarization of basin and sill geometry together with estimates of freshwater influx and rate of RSL change could significantly advance our ability to quantitatively interpret isolation basin stratigraphies for RSL change. Moreover, isolation basins have traditionally been viewed as ideal, quiet-water depositional environments that can record gradual changes in RSL. However, as we highlight in this review, isolation basins can act as excellent sedimentary archives for abrupt events, be they tsunami as illustrated by the early-Holocene Storegga event, landslide-triggered waves such as that recently observed in Disko Bugt, or large waves generated by ice-berg roll. Such events have potential to impact coastal communities in Greenland and other high latitude settings.

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Figure captions

Figure 1 (A) Schematic diagram illustrating the isolation process and accompanying isolation contacts. Modified from Kjemperud (1986). LAT is Lowest Astronomical Tide level, HAT is highest Astronomical Tide level. (B) Sample core of an isolation basin sequence collected from basin T4 (core 2a), part of the Upernivik isolation basin RSL record from close to Jakobshavn Isbrae in Disko Bugt, West Greenland. Core diameter is 5 cm. The isolation contact at 922-923 cm was dated to 6760±40 BP (7676–7513 cal yr BP, Beta-178165) (Long et al., 2006). (C) A well-exposed bedrock sill separating a contemporary tidal basin at Pâkitsoq (Disko Bugt) from the open coast. (D) A staircase of isolation basins from Arveprinsens Ejland (Disko Bugt) (Long et al., 1999).

Figure 2 A map of the Holocene marine limit from Greenland. Modified from Funder and Hansen (1996) and Weidick and Bennike (2007).

Figure 3 Location of places mentioned in the text. (A) Greenland. (B) Disko Bugt. (C) The location of four isolation basins described in the text from Kangerluarsuk (Disko Fjord).

Figure 4 The four isolation basins studied at Kangerluarsuk (Disko Fjord) showing core locations and accompanying lithostratigraphy. Altitudes are in meters above mean sea-level.

Figure 5 Diatom diagrams from each of the Kangerluarsuk (Disko Fjord) isolation basins. (A) D4, (B) D3, (C) D2, (D) D1. Stratigraphic notation: L = silt sand, hash = gyttja.

Figure 6 (A) Relative sea-level plot of the isolation basin data from Kangerluarsuk (Disko Fjord) obtained from basins D4-D1. Also plotted are the data from Eqalunnguit (K-3505) (Foged, 1989) and Qivittut (K-5576) (Bennike, 1995). (B) Isolation basin RSL data from other studies in Disko Bugt. Note that the data from Akulliit/Nuuk, Upernivik and Pâkitsoq are combined (Long et al., 2006) but different symbols are used to recognise that there may be some differential RSL change between these three sites. Age and altitude errors are detailed in Table 1.

Figure 7 Minimum ages for the timing of when local ice free conditions were established at sea level in Disko Bugt, based on isolation basin and other RSL and sea bed sediment cores from Disko Bugt (updated and revised from Long et al. (2003) and Weidick and Bennike (2007)). Ages are listed as the mid point of the two sigma calibrated age range in k cal yr BP.

Figure 8 Isolation basin RSL data from sites in Greenland: Outer Disko Bugt (Long et al. 2003, Long and Roberts, 2003), Sisimiut, central West Greenland (Long et al. (2009) and Bennike et al. (2011), Nanortalik, south Greenland (Sparrenbom et al., 2006a) and Ammassalik, southeast Greenland (Long et al., 2008).

Figure 9 Photographs illustrating a variety of isolation basin sills. (A) A tidal pond from close to Paamiut with (B) a sea-weed covered sill connecting the basin to the open sea exposed at low tide. (C) The sill that isolated basin N30 from the sea at Nanortalik, south Greenland (Sparrenbom et al., 2006a), with people standing on two potential sills. (D) A well-exposed bedrock sill from an isolation basin at Paamiut, southwest Greenland.

Figure 10 Results from a recent ice sheet modelling exercise by Simpson et al. (2009) (A) Modelled ice margin chronology for the Huy2 model; pink – 16 ka BP, dark blue – 14 ka BP, light blue – 12 ka BP, yellow – 10 ka BP, orange – 9 ka BP, red – 6 ka BP, green – 4 ka BP (minimum extent) and black – present-day. Modelled present day coastline is shaded dark grey. (B) RSL predictions for the Huy1 (light grey shaded area) and Huy2 (area enclosed by black dashed lines) ice models for the Arveprinsen Ejland area of Disko Bugt, based on a suite of 108 Earth viscosity profiles. The solid dark grey line denotes the RSL prediction for the best-fit Earth model of Huy1. The solid black line denotes the RSL prediction for the best-fit Earth model of Huy2. The RSL data is from isolation basins on Arveprinsen Ejland described in Table 1.

Table 1 List of radiocarbon dates from the main isolation basin studies referred to in the text from Greenland. The table is not an exhaustive record of all isolation basin records from Greenland but lists the majority of systematic studies conducted in the last 15 years. Radiocarbon dates are listed with a 2 sigma calibrated age range.

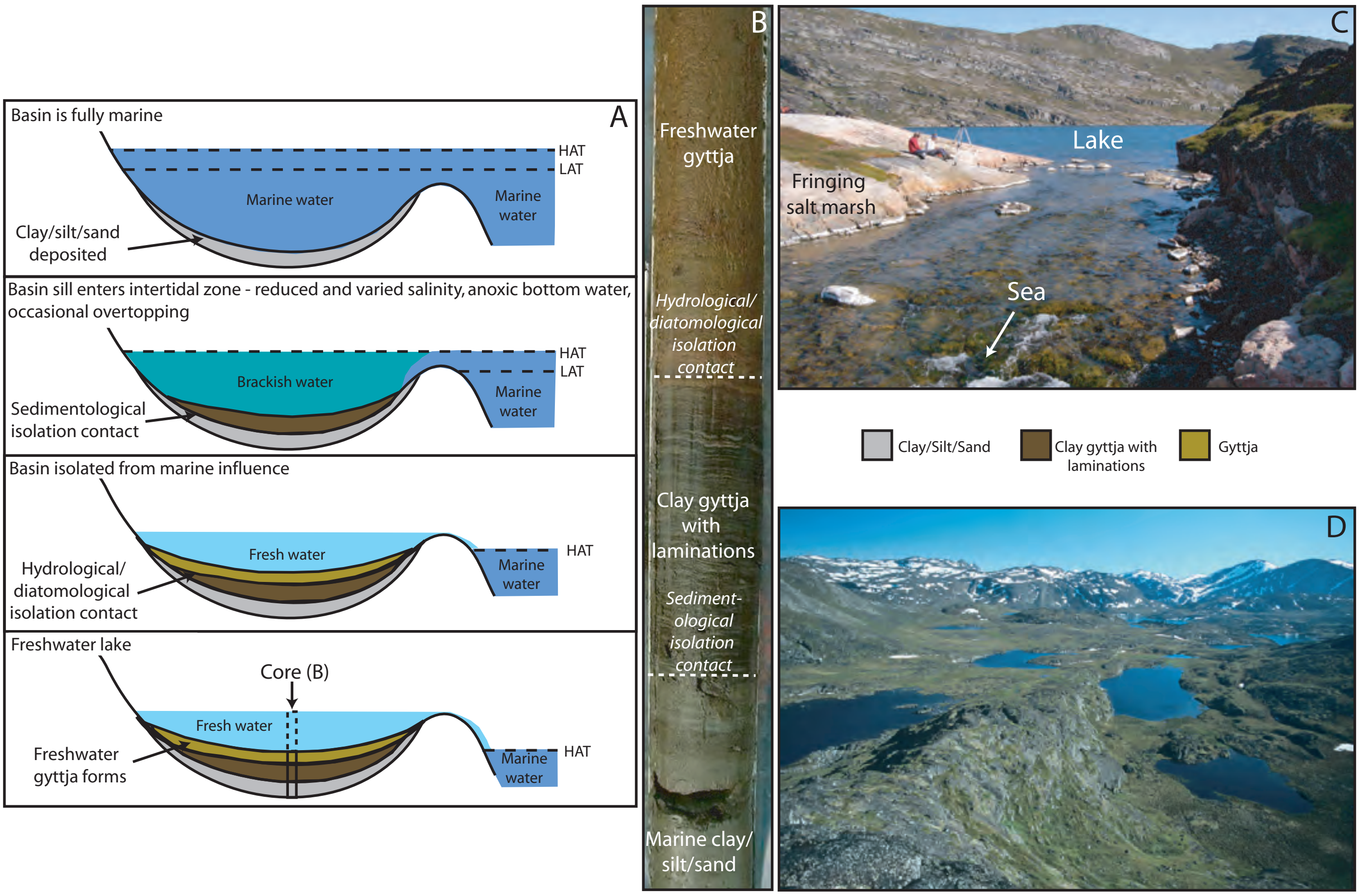
	Sill height (m MTL)	**Reference water level**	**Upper error +/-**	**Lower error +/-**	**RSL**	**Max cal age BP**	**Min cal age BP**	**Mid cal age BP**	**cal age error +/-**	**Mid cal age AD**	**^{14}C age**	**^{14}C +/- error**	**Lab code**
Disko Bugt													
Innaarsuit													
IV6 (Long et al., 2003)	80.67	isolation MHWST to HAT	0.36	0.36	79.61	10149	9553	9851	298	-7845	8780	80	AA39654
IV5 (Long et al., 2003)	32.99	isolation MHWST to HAT	0.52	0.52	31.96	7568	7273	7420.5	147.5	-5414.5	6530	70	AA39653
IV4 (Long et al., 2003)	21.23	isolation MHWST to HAT	0.59	0.59	20.28	6394	5999	6196.5	197.5	-4190.5	5440	60	AA39652
IV3 (Long et al., 2003)	2.91	isolation MHWST to HAT	0.48	0.48	1.98	4145	3729	3932	213	-1926	3620	60	AA39650
IV2 (Long et al., 2003)	-2.62	isolation MHWST to HAT	0.48	0.48	-2.62	2711	2212	2461.5	249.5	-455.5	2380	60	AA39649
IV2 (Long et al., 2003)	-2.62	ingression MHWST to HAT	0.48	0.48	-2.62	1292	1015	1153.5	138.5	852.5	1260	60	AA39648
Qeqertarsuatsiaq													
Q6 (Long and Roberts, 2003)	127.74	above marine limit	0.59	0.59	127.74	11041	10239	10640	401	-8634	9330	99	AA38842
U2 (Long and Roberts, 2003)	113.29	above marine limit	0.46	0.46	113.29	10547	10218	10382.5	164.5	-8376.5	9185	62	AA38844
Q5 (Long and Roberts, 2003)	98.14	isolation MHWST to HAT	0.56	0.56	97.09	10668	10241	10454.5	213.5	-8448.5	9270	61	AA38841
Q4 (Long and Roberts, 2003)	51.75	isolation MHWST to HAT	0.73	0.73	50.7	9272	8997	9134.5	137.5	-7128.5	8136	57	AA38840
U1 (Long and Roberts, 2003)	32.23	isolation MHWST to HAT	0.54	0.54	31.18	7647	7431	7539	108	-5533	6662	54	AA38843
Q3 (Long and Roberts, 2003)	8.46	isolation MHWST to HAT	0.5	0.5	7.41	5259	4852	5055.5	203.5	-3049.5	4397	45	AA38839
Q2 (Long and Roberts, 2003)	2.94	isolation MHWST to HAT	0.6	0.6	1.89	4567	4301	4434	133	-2428	3988	43	AA38838
Arveprinsens Ejland													
V0 (Long et al., 1999)	69.5	isolation MHWST to HAT	0.46	0.46	68.45	10181	9563	9872	309	-7866	8820	100	Beta-107879

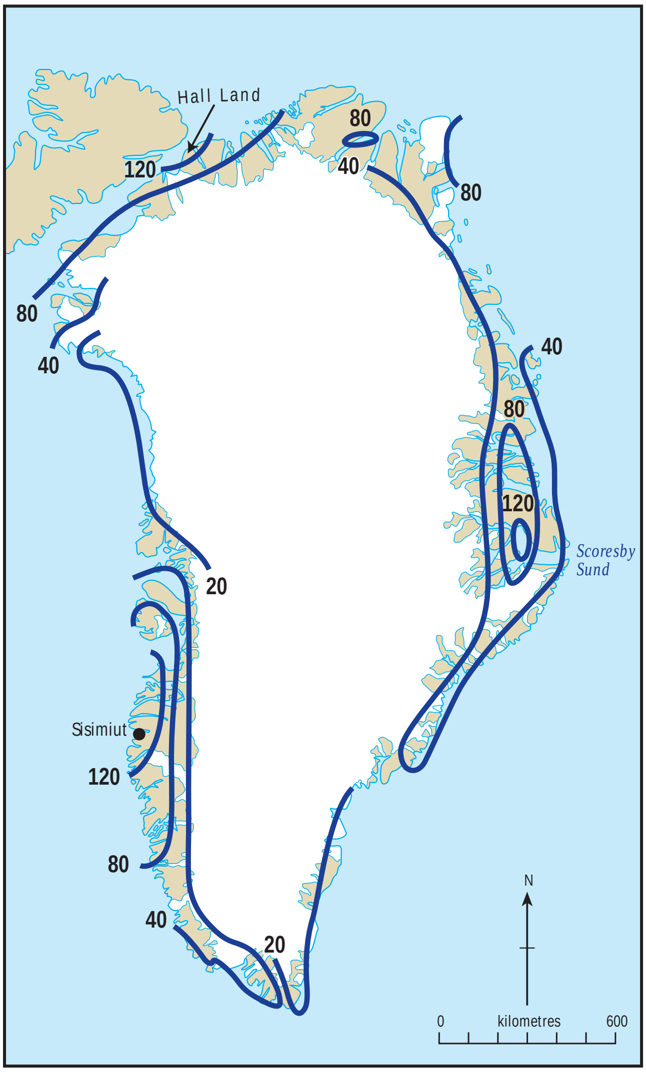
V2 (Long et al., 1999)	47.01	isolation HAT	0.38	0.38	45.91	8980	8387	8683.5	296.5	-6677.5	7780	120	Beta-112544
V5 (Long et al., 1999)	25.41	isolation MHWST to HAT	0.78	0.78	24.36	8151	7511	7831	320	-5825	6950	150	Beta-112543
V7 (Long et al., 1999)	18.16	isolation MHWST to HAT	0.25	0.25	17.11	7317	7156	7236.5	80.5	-5230.5	6290	40	Beta-110748
V8 (Long et al., 1999)	14.02	isolation MHWST to HAT	0.93	0.93	12.97	5310	4980	5145	165	-3139	4510	50	Beta-110747
V9 (Long et al., 1999)	-1.75	isolation MHWST	0.37	0.37	-2.75	4503	3721	4112	391	-2106	3750	130	Beta 110457
V9 (Long et al., 1999)	-1.75	ingression MHWST to HAT	0.37	0.37	-2.8	1407	1066	1236.5	170.5	769.5	1360	90	Beta 110456
Upernivik													
T7 (Long et al., 2006)	250	above marine limit	n/a	n/a	n/a	7823	7662	7742.5	80.5	-5736.5	6910	40	Beta-178170
T6 (Long et al., 2006)	175	above marine limit	n/a	n/a	n/a	7673	7512	7592.5	80.5	-5586.5	6750	40	Beta-178169
T5 (Long et al., 2006)	43.32	above marine limit	n/a	n/a	n/a	8996	8647	8821.5	174.5	-6815.5	7960	40	Beta-178168
T4 (Long et al., 2006)	31.56	isolation MHWST to HAT	0.28	0.28	30.59	7676	7513	7594.5	81.5	-5588.5	6760	40	Beta-178165
T3 (Long et al., 2006)	25.47	isolation MHWST to HAT	0.32	0.32	24.67	6895	6677	6786	109	-4780	5980	40	Beta-178171
T2 (Long et al., 2006)	11.93	isolation MHWST to HAT	0.35	0.35	11.15	5910	5715	5812.5	97.5	-3806.5	5060	40	Beta-178166
T1 (Long et al., 2006)	1.29	isolation MHWST to HAT	0.33	0.33	0.56	4826	4548	4687	139	-2681	4150	40	Beta-179167
T0 (Long et al., 2006)	-3.51	isolation MHWST to HAT	0.32	0.32	-4.38	3957	3693	3825	132	-1819	3540	40	Beta 178172
T0 (Long et al., 2006)	-3.51	ingression MHWST to HAT	0.32	0.32	-4.38	1812	1545	1678.5	133.5	327.5	1750	40	Beta 178173
Pâkitsoq													
P3 (Long et al., 2006)	41.21	above marine limit	n/a	n/a	n/a	7581	7270	7425.5	155.5	-5419.5	6814	38	KIA-23028
P1 (Long et al., 2006)	33.02	isolation MHWST to HAT	1	1	31.97	7429	7319	7374	55	-5368	6475	30	KIA-23026
P2 (Long et al., 2006)	36.08	above marine limit	n/a	n/a	n/a	7262	6680	6971	291	-4965	6113	121	KIA-23027
Akulliit/Nuuk													
AK4 (Long and Roberts, 2002)	89.52	above marine limit	n/a	n/a	n/a	9817	9436	9626.5	190.5	-7620.5	8585	86	AA39659
AK3 (Long and Roberts, 2002)	57.56	isolation MHWST	0.59	0.59	56.56	8647	8390	8518.5	128.5	-6512.5	7741	71	AA39658
AK2 (Long and Roberts, 2002)	36.89	isolation MHWST to HAT	0.22	0.22	35.84	7786	7571	7678.5	107.5	-5672.5	6812	61	AA39657
AK1 (Long and Roberts, 2002)	17.455	isolation MHWST to HAT	0.79	0.79	16.4	6651	6310	6480.5	170.5	-4474.5	5682	75	AA39656
N4 (Long and Roberts, 2002)	57.05	isolation (MHWST to HAT)	0.22	0.22	56	8376	8032	8204	172	-6198	7414	72	AA39664
N3 (Long and Roberts, 2002)	56.38	isolation (MHWST to HAT)	0.82	0.82	56.38	8604	8395	8499.5	104.5	-6493.5	7733	56	AA39665
N2 (Long and Roberts, 2002)	44.67	isolation (MHWST to HAT)	0.69	0.69	43.62	7998	7742	7870	128	-5864	7059	62	AA39661
N1 (Long and Roberts, 2002)	34.27	isolation (MHWST)	0.57	0.57	33.27	7563	7266	7414.5	148.5	-5408.5	6490	75	AA39660

Kangerluarsuk (Disko Island)													
D4 (this paper)	80.77	isolation (MHWST to HAT)	0.28	0.28	79.65	11265	10734	10999.5	265.5	-8993.5	9700	100	Beta-183253
D3 (this paper)	27.81	isolation (MHWST to HAT)	0.28	0.28	26.69	8537	8323	8430	107	-6424	7590	50	Beta-183252
D2 (this paper)	14.68	isolation (MHWST to HAT)	0.28	0.28	13.56	6439	6295	6367	72	-4361	5580	40	Beta-183251
D1 (this paper)	3.18	isolation (MHWST to HAT)	0.28	0.28	2.06	4409	4096	4252.5	156.5	-2246.5	3830	40	Beta-183250
Qivittut (Bennike, 1995)	16.5	isolation (HAT)	0.5	0.5	13	5716	5056	5386	330	-3380	4730	115	K-5776
Eqalunnguit (Foged, 1989)	22.5	isolation (HAT)	2.5	2.5	22.5	7819	7432	7625.5	193.5	-5619.5	6750	105	K-3505
Central west Greenland													
Sisimiut													
Sisi 15 (Bennike et al., 2011)	132.13	isolation (HAT)	0.5	0.5	129.9	10271	10115	10193	78	-8187	9020	60	LuS-8462
Sisi 11 (Bennike et al., 2011)	97.13	isolation (HAT)	0.5	0.5	94.9	10587	10130	10358.5	228.5	-8352.5	9070 +/-70 9270 +/-60		KIA-35512 LuS-8461
Sisi 10 (Bennike et al., 2011)	100.63	isolation (HAT)	0.5	0.5	98.4	10169	8762	9465.5	703.5	-7459.5	8170 +/-120 8630 +/-150 8485 +/-60		KIA-35509 KIA-35510 LuS-8460
Sisi 1 (Bennike et al., 2011)	42.03	isolation (HAT)	0.5	0.5	39.8	8524	8322	8423	101	-6417	7595	55	LuS-8459
Sisi 17 (Bennike et al., 2011)	20.63	isolation (HAT)	0.5	0.5	18.4	5283	4868	5075.5	207.5	-3069.5	4430	50	LuS-8783
NAG 15 (Long et al., 2009)	15.33	isolation	0.075	0.075	13	5881	5606	5743.5	137.5	-3737.5	4975	38	SUERC-15780
NAG 10 (Long et al., 2009)	10.01	isolation	0.16	0.16	7.68	4817	4449	4633	184	-2627	4102	38	SUERC-15783
NAG 2 (Long et al., 2009)	2.67	isolation	0.41	0.41	0.69	3578	3399	3488.5	89.5	-1482.5	3268	37	SUERC-17228
NAG 0 (Long et al., 2009)	0.14	isolation	0.31	0.31	-2.19	2343	2153	2248	95	-242	2246	37	SUERC-15785
NAG -1 (Long et al., 2009)	-1.89	isolation	0.28	0.28	-4.22	1345	1183	1264	81	742	1363	37	SUERC-15787
NAG -2 (Long et al., 2009)	-2.05	isolation	0.41	0.41	-4.025	2323	2120	2222.5	102.5	-216.5	2190	35	SUERC-15789
NAG -2 (Long et al., 2009)	-2.05	ingression	0.41	0.41	-4.025	2305	2002	2153.5	151.5	-147.5	2144	37	SUERC-15790
NAG 0 (Long et al., 2009)	0.14	ingression	0.31	0.31	-1.835	1281	1081	1181	100	825	1258	35	SUERC-15786
NAG -1 (Long et al., 2009)	-1.89	ingression	0.41	0.41	-3.865	1486	1300	1393	93	613	1480	37	SUERC-15788
Southern Greenland													
Qaqortoq													

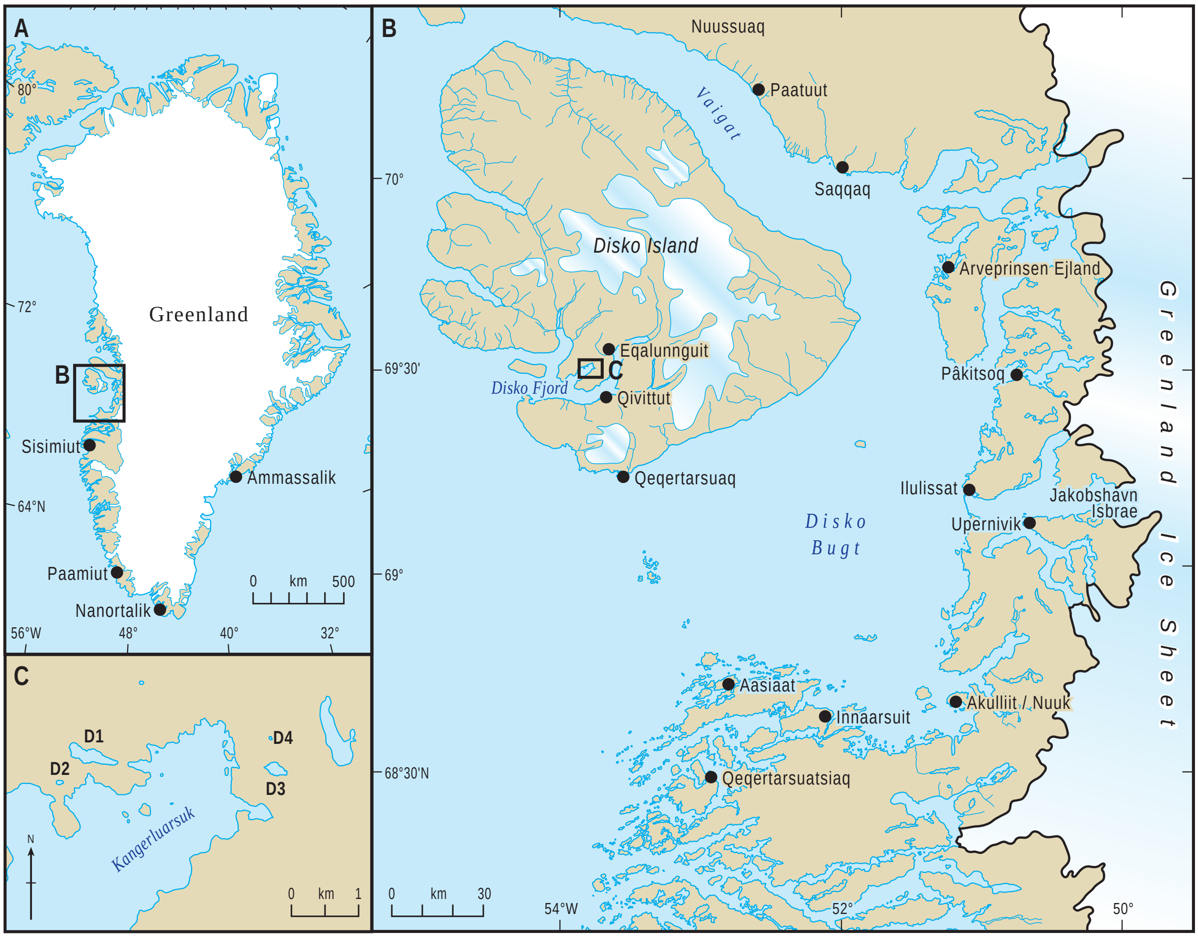
Q9 (Sparrenbom et al., 2006a)	32.7	isolation (HAT)	0.7	0.7	31.19	11200	10760	10980	220	-8974	9640	70	LuS-5877
Q5 (Sparrenbom et al., 2006a)	24	isolation (HAT)	0.7	0.7	22.49	10230	9790	10010	220	-8004	8930	60	LuS-5875
Q4 (Sparrenbom et al., 2006a)	14	isolation (HAT)	0.6	0.6	12.49	9500	8750	9125	375	-7119	8215	110	LuS-5873
K3 (Sparrenbom et al., 2006a)	8.9	isolation (HAT)	0.6	0.6	7.39	9610	9140	9400	260	-7394	8404 +/- 80 8505 +/- 70		LuS-5910 & LuS-5911
K1 (Sparrenbom et al., 2006a)	5.6	isolation (HAT)	0.5	0.5	4.09	9430	9030	9230	200	-7224	8255	70	LuS-5960
Q1 (Sparrenbom et al., 2006a)	0.7	isolation (HAT)	0.5	0.5	-0.81	9000	8340	8600	400	-6594	7835 +/-60 7625 +/-60 7955 +/-70		LuS-5927 LuS-5914 LuS-5926
Q3 (Sparrenbom et al., 2006a)	-1	isolation (HAT)	0.5	0.5	-2.51	9250	8600	8800	200	-6794	8060 +/-70 7945 +/-70		LuS-5918 LuS-5925
Q1 (Sparrenbom et al., 2006a)	0.7	ingression (HAT)	0.5	0.5	-0.81	3210	2870	3040	170	-1034	2890	50	LuS-5913
Q3 (Sparrenbom et al., 2006a)	-1	ingression (HAT)	0.5	0.5	-2.51	3880	3610	3745	135	-1739	3470	50	LuS-5923
Nu1 Nunataq (Sparrenbom et al., 2006a)	11.4	isolation (HAT)	0.6	0.6	9.89	11410	10790	11250	460	-9244	9700 +/-60 9845 +/-60		LuS-5920 & LuS-5878
Nanortalik													
N30 (Sparrenbom et al., 2006b)	1.25	ingression (HAT)	0.6	0.6	-0.26	1165.5	738.5	952	213.5	1054	1020	80	LuA-5312
N27 (Sparrenbom et al., 2006b)	-2.3	ingression (HAT)	0.5	0.5	-2.3	3000	1200	2100	900	-94	1550+-30 2810+-30		Poz-7172 & Poz-7173
N29 (Sparrenbom et al., 2006b)	-1.45	ingression (HAT)	1.1	2.1	-2.96	8140	2300	5220	2920	-3214	2290 +/-30 2395 +/-85 7300 +/-95		Poz-7176 LuA-5849 LuA-5316
N25 (Sparrenbom et al., 2006b)	-4.5	ingression (HAT)	0.6	0.6	-6.01	2740	2340	2540	200	-534	2450	85	LuA-5846
N28 (Sparrenbom et al., 2006b)	-5.95	ingression (HAT)	1.1	2.1	-7.46	5350	4650	5000	350	-2994	4375	95	LuA-5351
N29 (Sparrenbom et al., 2006b)	-1.45	isolation (HAT)	1.1	2.1	-2.96	9300	8600	8950	350	-6944	8085	95	LuA-5315
N25 (Sparrenbom et al., 2006b)	-4.5	isolation (HAT)	0.6	0.6	-6.01	9630	9470	9550	80	-7544	8560	50	Poz-7320
N28 (Sparrenbom et al., 2006b)	-5.95	isolation (HAT)	1.1	2.1	-7.46	9500	9030	9265	235	-7259	8250+-50 8310+-95		Poz-7178 & LuA-5350
N27 (Sparrenbom et al., 2006b)	-2.3	isolation (HAT)	0.5	0.5	-3.81	9600	9000	9300	300	-7294	8410	115	LuA-5348
N30 (Sparrenbom et al., 2006b)	-0.09	isolation (HAT)	0.6	0.6	-1.6	9530	9130	9330	200	-7324	8350	80	LuA-5311
N26 (Sparrenbom et al., 2006b)	5	isolation (HAT)	0.5	0.5	3.49	10650	9550	10100	550	-8094	8845+-115, 9110+-110, 9130+-105		LuA-5225 LuA-5226 LuA-5227
N21 (Bennike et al., 2002)	3	isolation (HAT)	2	2	1.49	10240	9750	9995	245	-7989	8930	80	Ua-15418
N22 (Bennike et al., 2002)	4	isolation (HAT)	2	2	2.49	10250	9700	9975	275	-7969	8905	90	Ua-15419
N16 (Bennike et al., 2002)	11	isolation (HAT)	2	2	9.49	10600	10210	10405	195	-8399	9240	95	Ua-15412
N19 (Bennike et al., 2002)	24	isolation (HAT)	2	2	22.49	12050	10550	11300	750	-9294	9810	175	Ua-15417
N24 (Bennike et al., 2002)	26	isolation (HAT)	2	2	24.49	12150	11150	11650	500	-9644	10015	120	Ua-15902
N18 (Bennike et al., 2002)	30	isolation (HAT)	2	2	28.49	12650	11250	11950	700	-9944	10200	110	Ua-15414
N14 (Bennike et al., 2002)	33	isolation (HAT)	2	2	31.49	15150	14150	14650	500	-12644	11665	125	Ua-14844
Tasiusaq													

Ta1 (Randsalu, 2007)	24.84	isolation (HAT)	0.5	0.5	23.33	8550	8210	8380	170	-6374	7590	60	LuS-7498
Ta3 (Randsalu, 2007)	10.93	isolation (HAT)	0.6	0.6	9.42	9270	8770	9020	250	-7014	8105	60	LuS-7500
Ta4 (Fredh, 2008)	-2.51	isolation (HAT)	0.6	0.6	-4.02	7440	7170	7305	135	-5299	7305	60	LuS-7521
Ta4 (Fredh, 2008)	-2.51	ingression (HAT)	0.6	0.6	-4.02	1340	1080	1210	130	796	1320	55	LuS-7524
East Greenland													
Ammassalik													
QT97 (Long et al., 2008)	97	above marine limit	n/a	n/a	n/a	11205	10775	10990	215	-8984	9659	67	SUERC-9428
QT93 (Long et al., 2008)	93	above marine limit	n/a	n/a	n/a	10766	10305	10535.5	230.5	-8529.5	9376	65	SUERC-9426
QT23 (Long et al., 2008)	23	isolation (MHWST to HAT)	0.64	0.64	21.13	9564	9411	9487.5	76.5	-7481.5	8523	67	SUERC-9425
QT11 (Long et al., 2008)	11.92	isolation (MHWST to HAT)	0.29	0.29	9.85	8538	8324	8431	107	-6425	7594	53	SUERC-9424
QT7 (Long et al., 2008)	7.52	isolation (MHWST to HAT)	0.34	0.34	5.65	8160	7942	8051	109	-6045	7202	51	SUERC-9421
QT3 (Long et al., 2008)	3.12	isolation (MHWST to HAT)	0.3	0.3	1.25	6791	6558	6674.5	116.5	-4668.5	5868	46	SUERC-9420

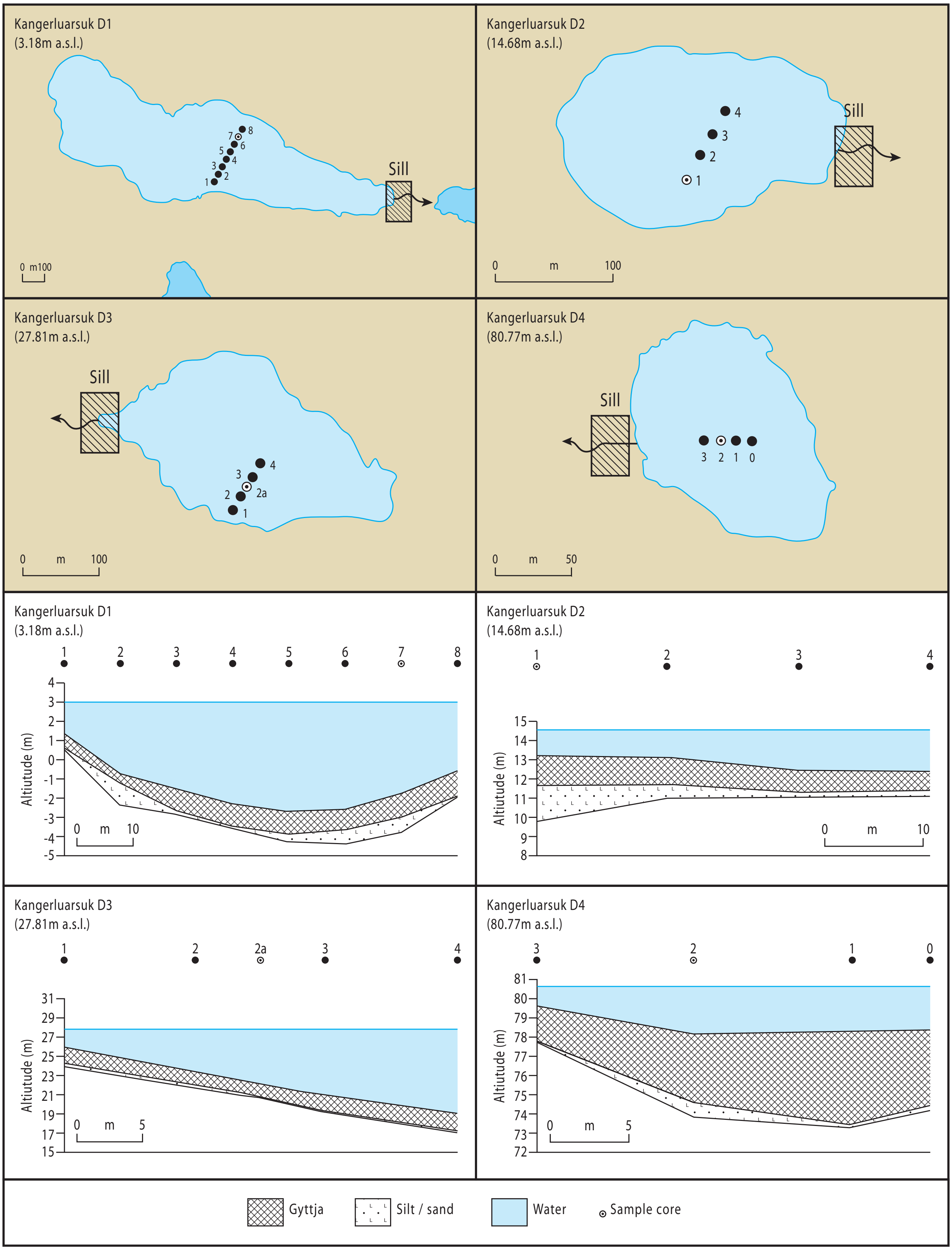

A
Basin is fully marine
HAT
LAT
Marine water
Marine water
Clay/silt/sand deposited
Basin sill enters intertidal zone - reduced and varied salinity, anoxic bottom water, occasional overtopping
HAT
LAT
Brackish water
Marine water
Sedimentological isolation contact
Basin isolated from marine influence
Fresh water
HAT
Marine water
Hydrological/ diatomological isolation contact
Freshwater lake
Core (B)
Fresh water
HAT
Marine water
Freshwater gyttja forms
B
Freshwater gyttja
Hydrological/ diatomological isolation contact
Clay gyttja with laminations
Sediment-ological isolation contact
Marine clay/ silt/sand
C
Lake
Fringing salt marsh
Sea
Clay/Silt/Sand
Clay gyttja with laminations
Gyttja
D

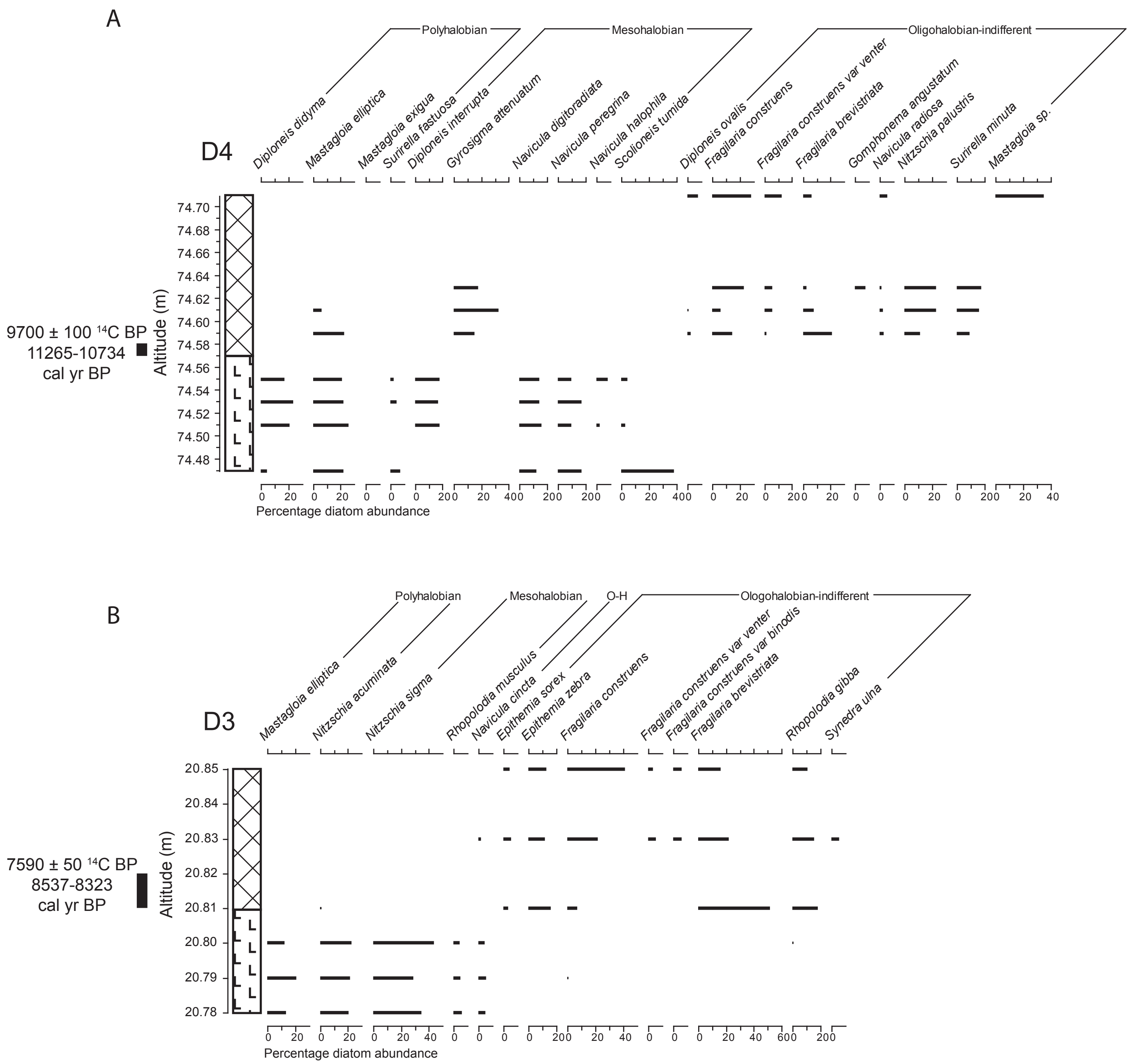


Hall Land
120
80
40
80
40
80
80
40
40
80
120
Scoresby Sund
20
Sisimiut
120
80
40
20
N
0
kilometres
600



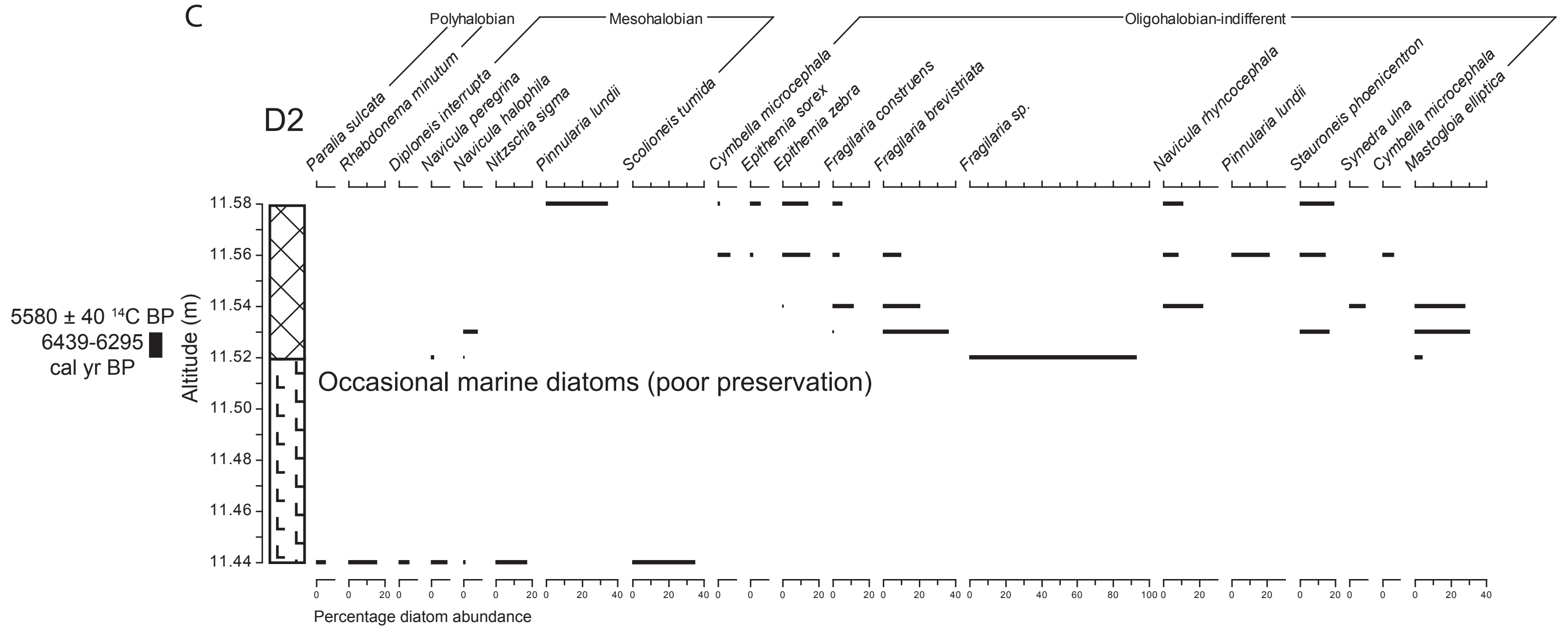
A
80°
72°
Greenland
B
Sisimiut
Ammassalik
64°N
Paamiut
Nanortalik
0 km 500
56°W 48° 40° 32°
B
Nuussuaq
Paatuut
Vaigat
Saqqaq
70°
Disko Island
Arveprinsen Ejland
Eqalunnguit
69°30'
C
Disko Fjord
Qivittut
Påkitsoq
Qeqertarsuaq
Ilulissat
Jakobshavn Isbrae
Disko Bugt
Upernivik
69°
Greenland Ice Sheet
Aasiaat
Innaarsuit
Akulliit / Nuuk
68°30'N
Qeqertarsuatsiaq
0 km 30
54°W 52° 50°
C
D1
D2
D4
D3
Kangerluarsuk
N
0 km 1


Kangerluarsuk D1
(3.18m a.s.l.)
Sill
0 m100
Kangerluarsuk D2
(14.68m a.s.l.)
Sill
0 m 100
Kangerluarsuk D3
(27.81m a.s.l.)
Sill
0 m 100
Kangerluarsuk D4
(80.77m a.s.l.)
Sill
0 m 50
Kangerluarsuk D1
(3.18m a.s.l.)
Altiutude (m)
0 m 10
Kangerluarsuk D2
(14.68m a.s.l.)
Altiutude (m)
0 m 10
Kangerluarsuk D3
(27.81m a.s.l.)
Altiutude (m)
0 m 5
Kangerluarsuk D4
(80.77m a.s.l.)
Altiutude (m)
0 m 5
Gyttja
Silt / sand
Water
Sample core



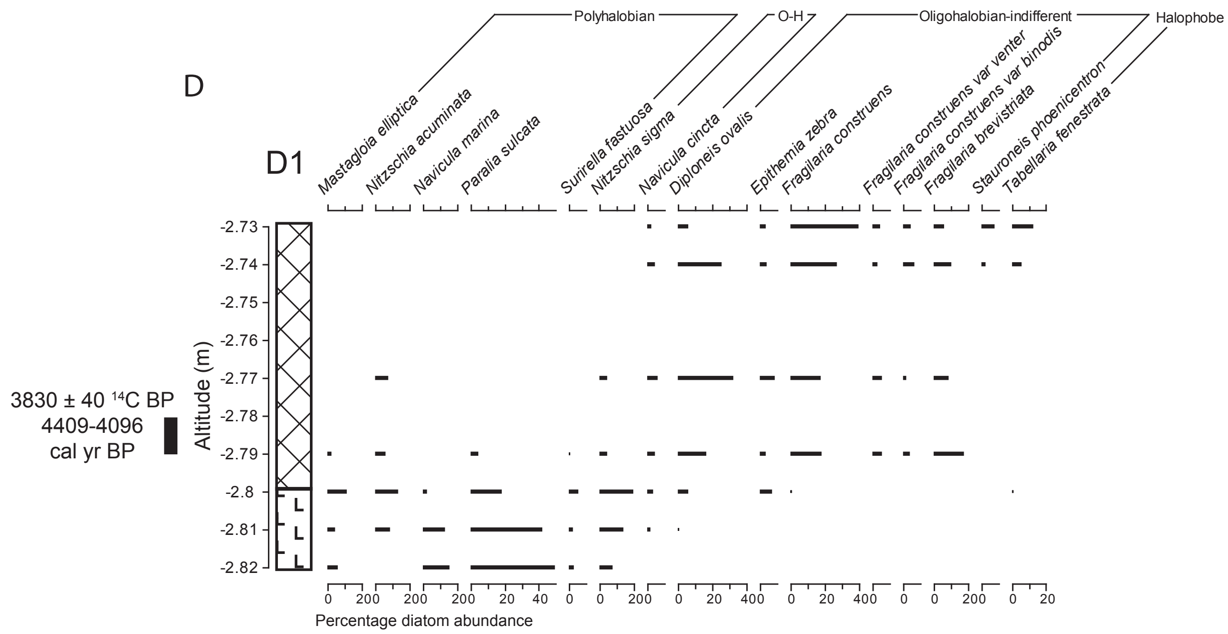
A
Polyhalobian
Mesohalobian
Oligohalobian-indifferent
D4
Diploneis didyma
Mastagloia elliptica
Mastagloia exigua
Surirella fastuosa
Diploneis interrupta
Gyrosigma attenuatum
Navicula digitoradiata
Navicula peregrina
Navicula halophila
Scolioneis tumida
Diploneis ovalis
Fragilaria construens
Fragilaria construens var venter
Fragilaria brevistriata
Gomphonema angustatum
Navicula radiosa
Nitzschia palustris
Surirella minuta
Mastagloia sp.
Altitude (m)
9700 ± 100 14C BP
11265-10734
cal yr BP
Percentage diatom abundance
B
Polyhalobian
Mesohalobian
O-H
Ologohalobian-indifferent
D3
Mastagloia elliptica
Nitzschia acuminata
Nitzschia sigma
Rhopolodia musculus
Navicula cincta
Epithemia sorex
Epithemia zebra
Fragilaria construens
Fragilaria construens var venter
Fragilaria construens var binodis
Fragilaria brevistriata
Rhopolodia gibba
Synedra ulna
Altitude (m)
7590 ± 50 14C BP
8537-8323
cal yr BP
Percentage diatom abundance

C

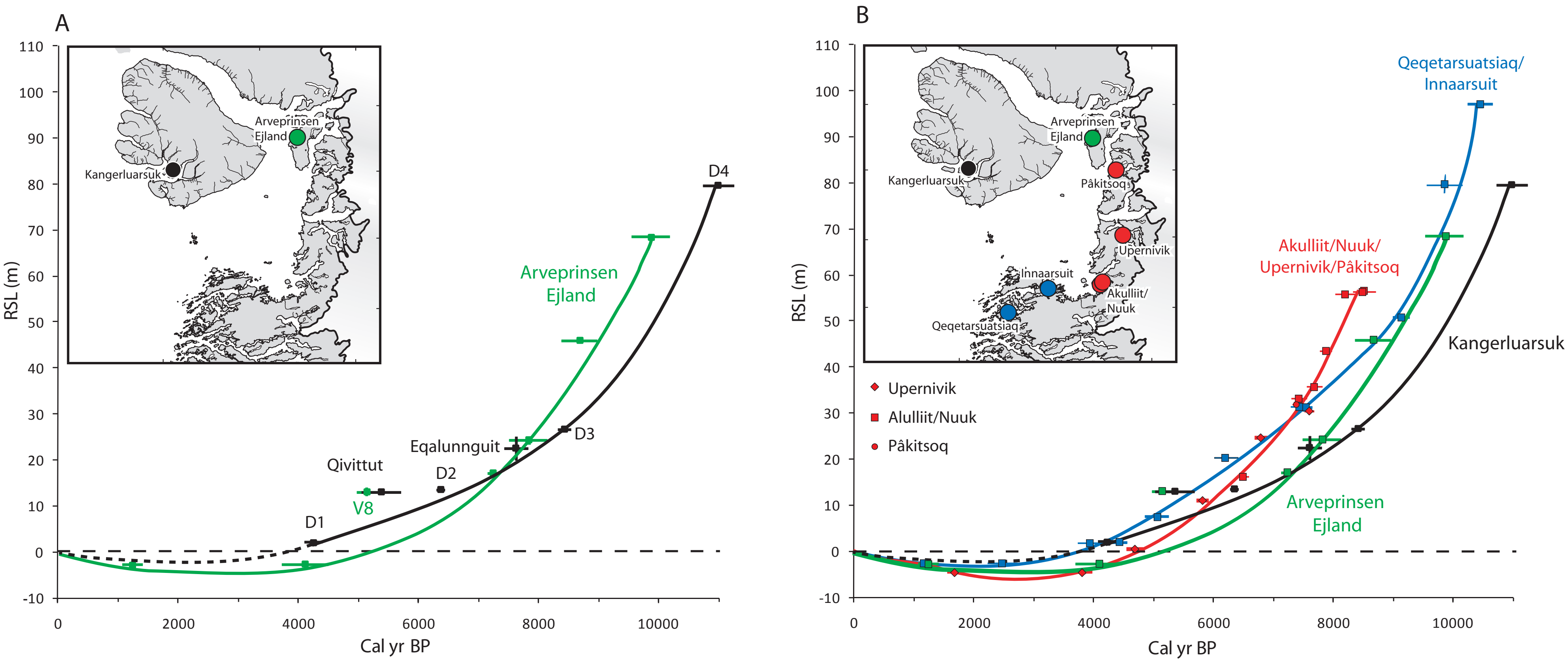


Polyhalobian
Mesohalobian
Oligohalobian-indifferent
D2
Paralia sulcata
Rhabdonema minutum
Diploneis interrupta
Navicula peregrina
Navicula halophila
Nitzschia sigma
Pinnularia lundii
Scolioneis tumida
Cymbella microcephala
Epithemia sorex
Epithemia zebra
Fragilaria construens
Fragilaria brevistriata
Fragilaria sp.
Navicula rhyncocephala
Pinnularia lundii
Stauroneis phoenicentron
Synedra ulna
Cymbella microcephala
Mastogloia elliptica
5580 ± 40 14C BP
6439-6295
cal yr BP
Altitude (m)
Occasional marine diatoms (poor preservation)
Percentage diatom abundance

D

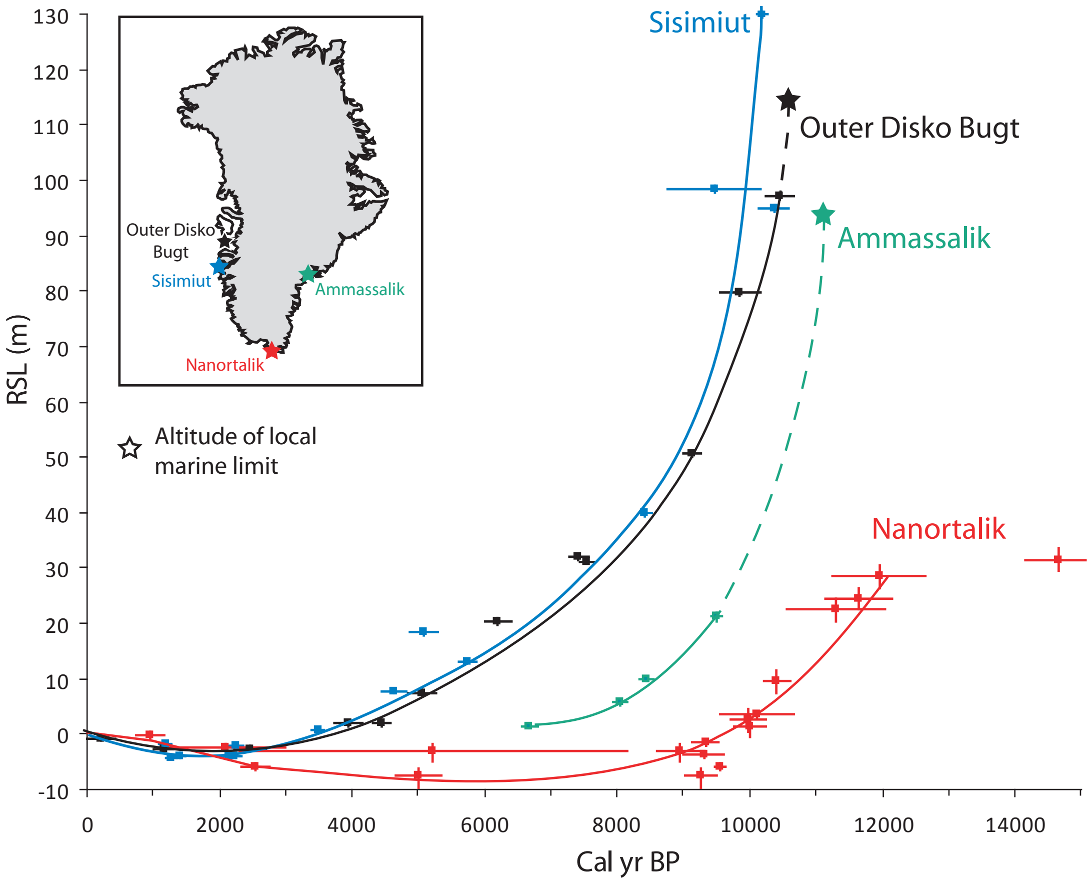


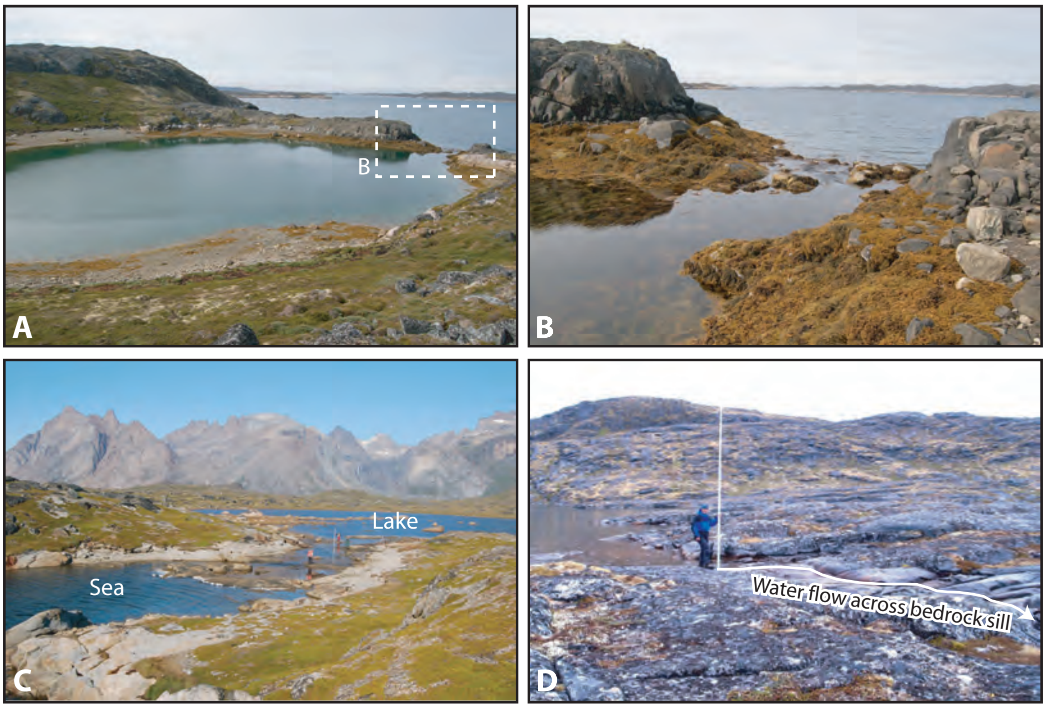
Polyhalobian
O-H
Oligohalobian-indifferent
Halophobe
D1
Mastagloia elliptica
Nitzschia acuminata
Navicula marina
Paralia sulcata
Surirella fastuosa
Nitzschia sigma
Navicula cincta
Diploneis ovalis
Epithemia zebra
Fragilaria construens
Fragilaria construens var venter
Fragilaria construens var binodis
Fragilaria brevistriata
Stauroneis phoenicentron
Tabellaria fenestrata
3830 ± 40 14C BP
4409-4096
cal yr BP
Altitude (m)
Percentage diatom abundance

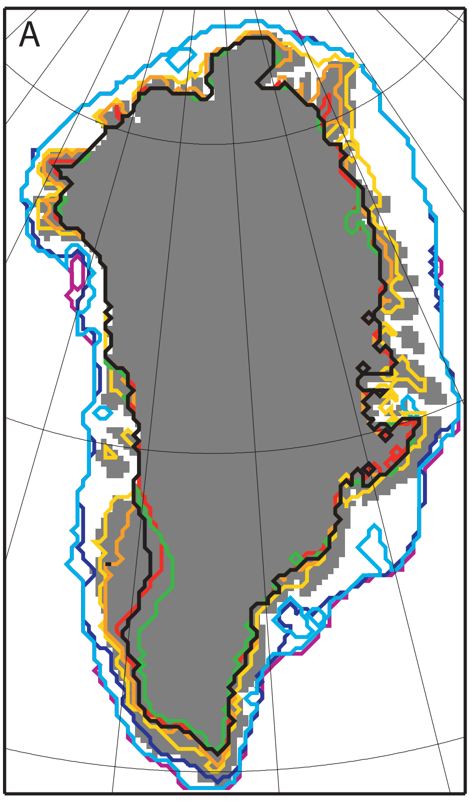


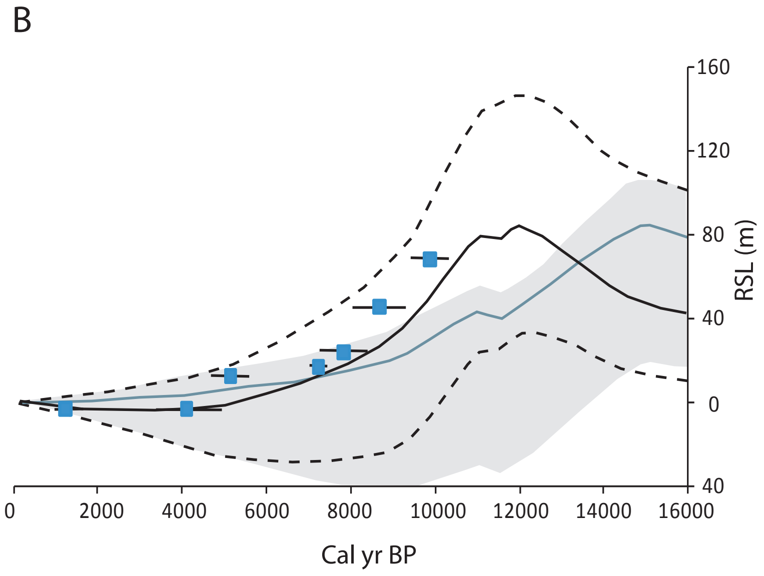
A
B
RSL (m)
Cal yr BP
Arveprinsen Ejland
Kangerluarsuk
Pâkitsoq
Upernivik
Innaarsuit
Akulliit/ Nuuk
Qeqetarsuatsiaq
D4
D3
D2
D1
Eqalunnguit
Qivittut
V8
Upernivik
Alulliit/Nuuk
Pâkitsoq
Qeqetarsuatsiaq/ Innaarsuit
Akulliit/Nuuk/ Upernivik/Pâkitsoq


12.4
11.8
11.6
11.0
10.2
10.4
10.3
Vaigat
10.0
70°
Disko Island
9.9
Arveprinsen Ejland
10.0
Greenland Ice Sheet
69°30'
11.0
Disko Fjord
Påkitsoq
7.7
10.3
Ilulissat
Jakobshavn Isbrae
6.1
10.5
Disko Bugt
9.8
Upernivik
8.4
69°
9.9
9.8
7.9
9.6
Aasiaat
10.0
Innaarsuit
Akulliit / Nuuk
68°30'N
10.5
Qeqertarsuatsiaq
0
km
30
54°W
52°
50°


Sisimiut
Outer Disko Bugt
Ammassalik
Nanortalik
Outer Disko Bugt
Sisimiut
Ammassalik
Nanortalik
Altitude of local marine limit
RSL (m)
Cal yr BP
130
120
110
100
90
80
70
60
50
40
30
20
10
0
-10
0
2000
4000
6000
8000
10000
12000
14000


A
B
B
C
Lake
Sea
D
Water flow across bedrock sill


A


B
160
120
80
40
0
40
RSL (m)
0
2000
4000
6000
8000
10000
12000
14000
16000
Cal yr BP